PEDOGENESIS AND EROSION HISTORY IN A HIGH RAINFALL, MOUNTAINOUS DRAINAGE BASIN-CROPP RIVER, NEW ZEALAND

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ABSTRACT

Cropp River drains a 28.5 km^2 drainage basin in the western Southern Alps. Extreme erosional modification of this formerly glaciated valley has occurred in response to high rainfall (10,800 mm a^{-1}) and rapid tectonic uplift (12 $\pm$ 2 mm a^{-1}). Studies of geomorphology, soil development and distribution, and contemporary forms of erosion provide the basis for interpreting the history of post-glacial erosion.

During late Pleistocene (14 to 22,000 years BP) glacial advances Cropp basin was largely occupied by ice. Till dated at 10,250 $\pm$ 150 years BP, in the middle reaches of Cropp River, provides a maximum age for the post-glacial topography of the upper basin. U-shaped cirques are present at the heads of the larger tributaries. Elsewhere the glacial valley form has been largely destroyed by fluvial and mass movement erosion. This has produced a dominantly erosional landscape with extremely steep slopes (40 to 70° common), intense dissection, and steep V-shaped stream channels.

Soil development sequences, independently dated by dendrochronology and radiocarbon dating, were examined on both gently and steeply sloping topography. Soil development is rapid and progressively forms recent soils (<200 years), yellow-brown earths (500 to 1000 years), podzolised yellow-brown earths (1000 to 1500 years) and gley podzols (2500 to 10,000 years). All soils are strongly leached indicated by rapid soil acidification, and the dominance of exchangeable Al and H even in recent soils. The older soils (>1000 years) are both podzolised and gleyed. Eluvial-illuvial coefficients, calculated from total element analyses, indicated the losses and gains of major elements during soil development.

Morphology of soils on steep slopes under both scrub-forest and grassland vegetation indicates most soils range in age from tens to a few thousand years. Soils are removed from steep slopes by episodic and progressive erosional processes, generally within 500 to 1000 years. Evidence of active erosion is widespread although largely concealed by dense vegetation cover.

Natural revegetation of eroded sites is rapid despite low soil nutrient status. On steep slopes eroded sites have complete vegetation cover within 50 years, and within 500 to 1000 years have vegetation similar to uneroded areas. Obvious visual evidence of erosion is short-lived because of rapid revegetation and plant succession.

Keywords: Erosion; soil stratigraphy; soil genesis; chronosequences; podzolisation; gleying; radiocarbon dates; glacial history; plant succession; subalpine scrub; Cropp River; Westland; Southern Alps; McKerrow soils.

FRONTISPIECE

General view of Cropp River from its confluence with the Whitcombe River (foreground) to the headwaters on Galena Ridge (centre top). Looking west with the Tasman Sea in the background. Noisy Creek is the hanging valley in the right centre of the photo.

CONTENTS

Page

Page

LIST OF FIGURES

LIST OF TABLES

CHAPTER 1 - BACKGROUND TO THE STUDY

1.1 INTRODUCTION

This study of soil erosion in a region of very high annual rainfall forms part of a multi-disciplinary programme of studies of small and large drainage basins in a transect across the Southern Alps in the region of the Rakaia and Hokitika Rivers (see Figs. 1 and 2 for localities mentioned in text). These studies, undertaken under the auspices of the National Water and Soil Conservation Authority (NWASCA) by the Ministry of Works and Development (MWD), aim to measure, and predict with precision, the ways in which mountain drainage basins are altered by land use, by the weather and tectonism, and to determine how these alterations are expressed in changes in water and sediment yield, in water quality, and in soil and vegetation cover (MWD, 1977a, pp 18-21).

Seven small drainage basins have been studied. Dry Acheron Stream (Basher, 1980; Basher *et al*, 1980; Whitehouse and McSaveney, 1980; Griffiths and Hicks, 1980) and Ryton River (Tonkin *et al*, 1981; Harrison, 1982) are representative of the drier eastern high country (1500 to 2000 mm annual precipitation), and are similar to previously studied areas at Torlesse and Camp Streams (Wilde, 1974; Hayward, 1980) and Porters Pass (Molloy 1963, 1964). Scree hydrology has been studied at Talus Tarn in the wetter (3255 mm a^{-1}) alpine regions of the upper Rakaia valley. Flows from the Ramsay and Lyell Glaciers in the upper Rakaia valley, and sediment yield into the lake at Lyell Glacier, have been monitored. This provides data on rates of erosion from the glaciated greywacke ranges on the Main Divide, with a precipitation of 4800 to 5400 mm a^{-1}. At Ivory Glacier, mass, energy and water balances were measured from 1969 to 1975 as part of the International Hydrological Decade representative basin program (Anderton and Chinn, 1978). Subsequently rainfall measurement has continued and sediment yield into a proglacial lake has been monitored (Robinson, 1981) to give estimates of rates of erosion from a glaciated basin with a precipitation of 9,200 mm a^{-1}. Cropp River, the seventh of the study sites, lies entirely within the zone of annual precipitation exceeding

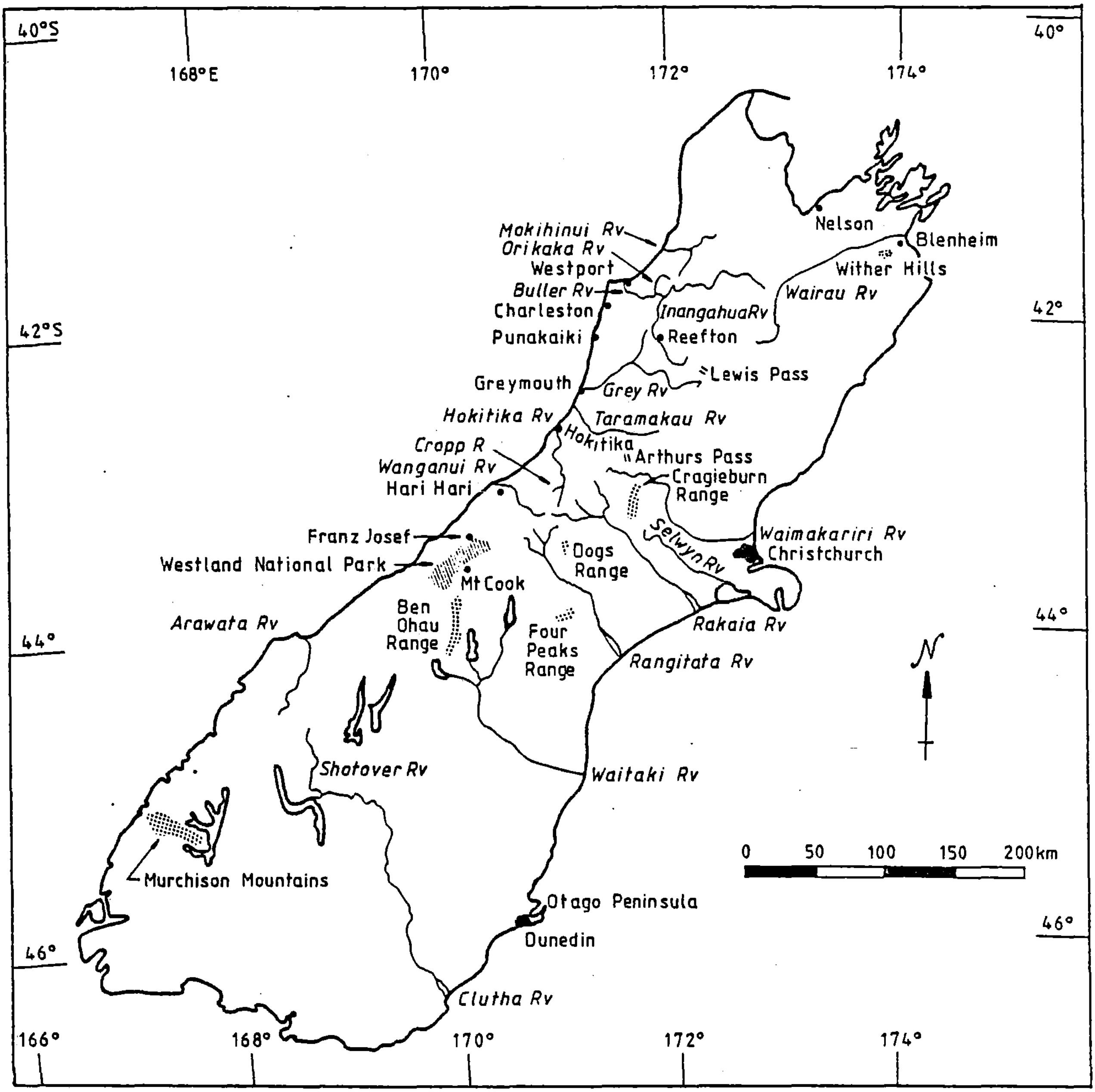


FIG. 1: Map of South Island, New Zealand showing localities mentioned in text.

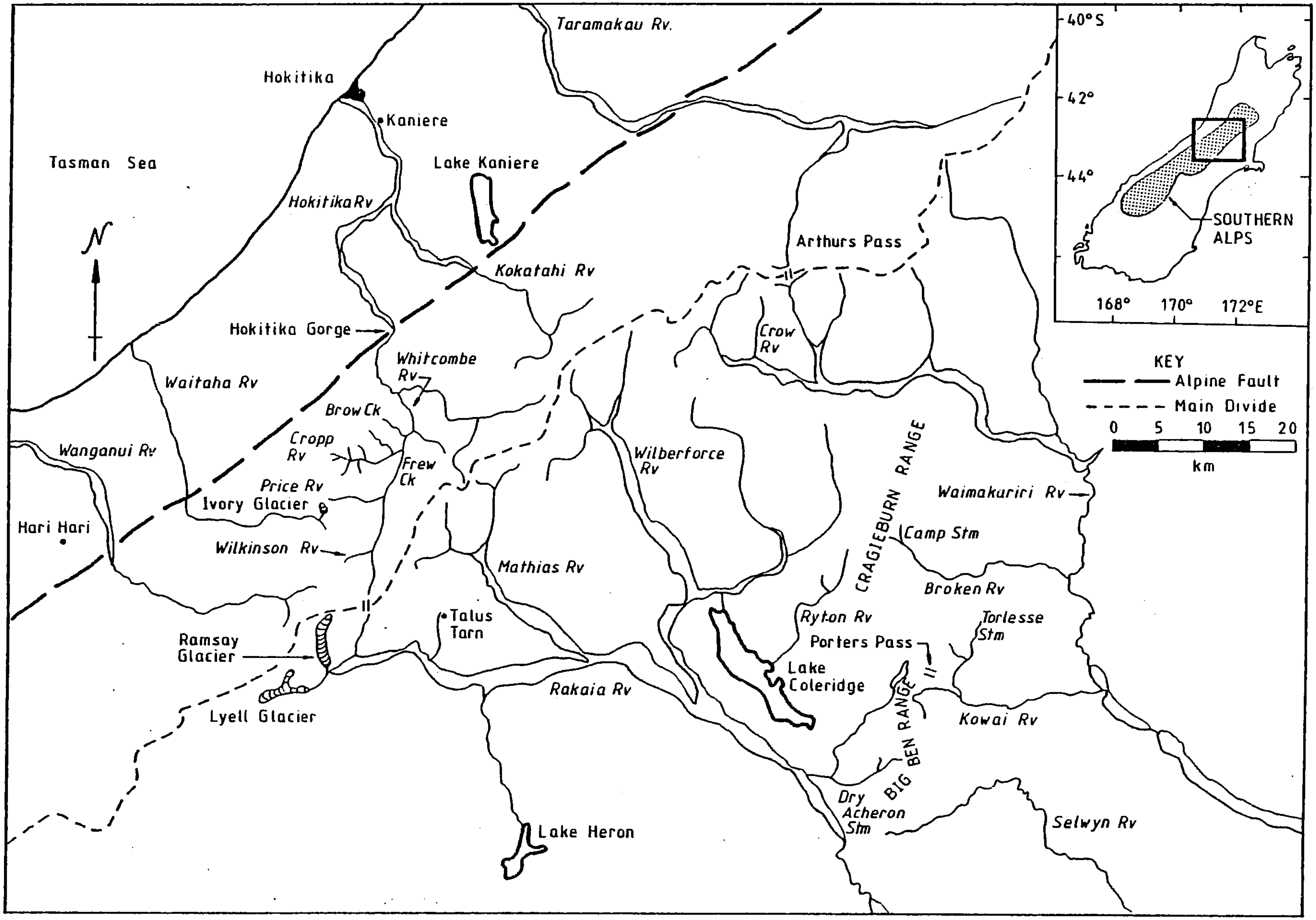


FIG. 2: Map of Rakaia-Hokitika area showing localities mentioned in text.

10,000 mm (Griffiths and McSaveney, 1983a). It is an example of the formerly glaciated areas west of the Main Divide that are subject to high annual precipitation and high rates of tectonic uplift.

The following studies have been reported for Cropp River:

(a) Hawkes (1981) describes bedrock and structural geology, uplift rates and the major geomorphological features;

(b) Norton (1983a) discusses factors influencing the population dynamics of subalpine *Libocedrus bidwillii* (kaikawaka or pahautea) forests. Norton (1983b) analyses the dendroclimatic potential of L. bidwillii growing at several sites in Cropp basin;

(c) Griffiths and McSaveney (1983a) describe the hydrology (precipitation, runoff and sediment yield) of Cropp River.

These studies and their relationship with the present study will be described in more detail in later chapters of this thesis.

1.2 PREVIOUS STUDIES OF EROSION IN THE SOUTH ISLAND MOUNTAIN LANDS

It is not intended to review all previous studies of rates and processes of erosion in the Southern Alps because:

(a) the published literature deals almost exclusively with areas east of the Main Divide; and

(b) a number of authors have recently presented reviews of various aspects of erosion research in, and changing attitudes towards, these mountain lands (Howard, 1978; Mosley, 1978; Pearce and O'Loughlin, 1978; Hayward, 1980; O'Connor, 1980a; Mather, 1982; O'Loughlin and Pearce, 1982; Whitehouse, 1984).

This section places studies of Cropp basin in the perspective of erosion research in the Southern Alps. It describes how attitudes have changed towards the assessment of soil erosion in view of recent scientific evidence, and how techniques of assessment are altering.

Early surveys of soil erosion in the Southern Alps (Zotov, 1938, 1940; Committee of Enquiry, 1939; Cumberland, 1944; Gibbs *et al*, 1945) emphasise vegetation depletion rather than soil erosion *per se*. Implicit in this approach was a belief that soil erosion was an inevitable consequence of vegetation depletion. Gibbs *et al* (1945) define soil (or accelerated) erosion as that occurring when natural vegetation is disturbed. Therefore the early surveys of soil erosion concentrate on the eastern mountain lands where deforestation, pastoral farming and introduced animals (rabbits, hares, deer, goats, chamois, opossums) had their greatest influence.

Interpretation of these early surveys led virtually all subsequent research to be undertaken in the eastern mountain lands. Much effort was directed towards revegetation as a means of mitigating erosion (O'Connor, 1962, 1976, 1980b; Dunbar and Adams, 1972; Dunbar *et al*, 1977; Molloy and Connor, 1970; Mugambi, 1971; Orwin, 1978; Ledgard and Miller, 1980; Baker and Ledgard, 1981; Davis, 1981a, b, c; Ledgard and Baker, 1982), while less effort went into measuring and understanding rates and processes of erosion. As Hayward (1980) points out, "Those who first advocated soil conservation saw a clear need for remedial action in preference to research."

From about 1970 greater effort was directed towards scientific study of rates and processes of soil erosion with the result that a number of long held beliefs were disputed. For example: it is now recognised that many screes are old features that have never been vegetated (Whitehouse *et al*, 1980; Whitehouse and McSaveney, 1983); the presumed direct relationship between hillslope erosion and downstream aggradation problems is erroneous (Griffiths, 1979a; Hayward, 1980; Mosley, 1980); and Hortonian overland flow is rare since rainfall intensities rarely exceed infiltration capacities of soils (Gillingham, 1964; Soons, 1970; Hayward, 1977). But even the recent review of O'Loughlin and Pearce (1982) notes that "in view of the widely held, and often expressed, view that erosion rates are extremely rapid, in at least some regions, relatively few studies of rates and processes of erosion have been undertaken, even in those areas where concern over supposedly high, man-induced erosion rates is greatest".

Recent research has considerably improved our understanding of rates and processes of erosion in the Southern Alps. Frost action and needle ice formation have been the subject of several studies (Gradwell, 1954, 1955, 1957, 1960; Soons, 1968, 1971; Soons and Rayner, 1968) because the resulting soil disturbance is a precursor to wind erosion (Butterfield, 1971) soil creep (Owens, 1969), and sheet erosion (Hayward, 1967, 1968, 1969). Whitehouse *et al* (1980) and Whitehouse and McSaveney (1983) describe how screes can be dated, their mechanism of formation, and show that many screes are very old features of the landscape. Rock avalanching as a geomorphic process in the Southern Alps is described by Whitehouse (1981, 1983) and Whitehouse and Griffiths (1983). Pierson (1980a, b, 1981) describes the mechanics and impact of debris flows as an erosion process.

Considerable attention has been given to the relationship between hillslope erosion and downstream sediment accumulations. Griffiths (1979a) provides evidence that aggradation in the lower Waimakariri River is not primarily due to sediment derived from headwater areas. Cuff (1974, 1977, 1981), Hayward (1980) and Mosley (1980) demonstrate that sediment transported in mountain drainage basins is largely obtained from riparian sources, most of which are large natural features. Often these sources are of previously eroded sediment that has been held in storage, sometimes since the late Pleistocene. The linkages between hillslope erosion, sediment transport and resultant channel characteristics in the upper Kowai Basin were shown to be complex, highly variable and strongly influenced by low frequency/large magnitude events (Ackroyd and Blakely, 1984; Beschta 1983a, b).

Regional and local rates of erosion have been estimated from sediment yield measurements, which integrate the products of many erosion processes, and are also affected by sediment storage (Pearce and O'Loughlin, 1978; Mosley, 1978). This indirect approach has been used because of concern about river sediment loads and because of the difficulties in making quantitative estimates of hillslope erosion. Adams (1978, 1980a), McSaveney (1978), Griffiths (1979b, 1981), Thompson and Adams (1979), Hayward (1980) and Griffiths and McSaveney (1983a) provide estimates for small and large catchments within the Southern Alps. Rainfall appears to be the most important variable determining sediment yield. Griffiths (1981) provides an empirical power law relationship between yield and rainfall for catchments underlain by various rock types:

$$G \propto P^{2.4}$$

where G = specific annual suspended sediment yield
and P = catchment mean annual rainfall

While this approach does not directly indicate rates of hillslope erosion, it does provide an index to the erosional activity of drainage basins and a basis for regional comparison (O'Loughlin and Pearce, 1982). The implication of the studies cited above is that as rainfall increases, sediment yield increases exponentially. In large basins, most of the sediment yield can thus be produced from small areas of the basin subject to high rainfall. The importance of this relationship has become obvious now that the distribution of precipitation across the Southern Alps (see fig. 10) has been quantified (McSaveney, 1978; Griffiths and McSaveney, 1983b).

The background of geological instability and the relationship of plate tectonics in the New Zealand region to regional rates of erosion has been reviewed by Ackroyd (1978), Adams (1978, 1980a), and Whitehouse (1985a). Adams (1978, 1980a) made an assessment of the relationship between uplift and erosion in the Southern Alps and proposed that tectonic uplift is approximately in balance with denudation. This is supported by the absence of any gravity anomaly associated with the Southern Alps which suggests they are in isostatic equilibrium (Eiby and Reilly, 1976). Whitehouse (1985a) describes the regional geomorphology of the central Southern Alps in relation to plate tectonics. Three geomorphological regions parallel the strike of the Alps. In the western region uplift, precipitation and erosion rates are greatest, and erosional landforms predominate. In the axial region elevation is greatest, although erosion and uplift rates are lower than in the west, and glacial landforms predominate. In the eastern region erosion and uplift rates are least and the basin and range topography contains many relict glacial and periglacial landforms. On a more local scale, many studies document the association of incompetent rock types, faulting and crush zones with the more dramatic expressions of erosion (Cuff, 1974, 1977, 1981; Blair, 1972; Main, 1975; Brommell *et al*, 1980 ; Hawkes *et al*, 1980; Mortimer and Whitehouse, 1980; Pierson, 1980a, b; Whitehouse and McSaveney, 1980; Whitehouse *et al*, 1982).

There are a number of detailed studies in which a pedological approach is used to assess rates and processes of soil erosion on hillslopes. The early work of Molloy (1962, 1963, 1964) at Porters Pass has been followed by studies in the Ben Ohau Range and Mt Cook area (Archer 1973, 1976, 1978; Archer *et al*, 1973; Archer and Cutler, 1983), Four Peaks Range (Ives, 1970; Ives and Cutler, 1972), Dogs Range (Harvey 1974), Cragieburn Range (Wilde, 1974; Tonkin *et al*, 1981; Harrison, 1982) and Big Ben Range (Basher, 1980). These studies use aspects of the soil pattern, soil properties and soil stratigraphy to elucidate the complex erosional and depositional history of the eastern Southern Alps. Recent reviews (Howard, 1978; O'Connor, 1980a; Whitehouse, 1984) discuss the complexity of inter-related factors that have determined the erosion mosaic in the eastern mountain lands. The authors review the impact of European man on the landscape in the perspective of the major climatic changes of the Pleistocene and Aranuian, and of the impact of pre-European vegetation modification.

In contrast to the relatively extensive literature on erosion in the eastern mountain lands, there are few studies of areas west of the Main Divide. O'Loughlin and Pearce (1982) in their review of erosion processes in the Southern Alps cite no references for areas in the western Southern Alps but suggest that "the zone of maximum slope instability coincides roughly with the zone of maximum precipitation and serious vegetation depletion". A number of studies are reported for lowland and hill country areas on the West Coast (O'Loughlin and Pearce, 1976; O'Loughlin *et al*, 1978, 1980, 1982; Laffan, 1979a, b; Mosley, 1982) but few are available for steep mountain catchments west of the Main Divide (Hawkes, 1981; Griffiths and McSaveney, 1983a). This partly reflects the fact that West Coast catchments were viewed by early workers as having lower rates of erosion than east coast catchments (Holloway, 1959) and were given low erosion severity rankings on Land Resource Inventory Worksheets (MWD, 1979). Some concern about the effect of introduced animals (especially deer and opossums) on vegetation condition led early authors to assert that farming on the lowland plains of the Hokitika River might have to be abandoned, as a result of increased erosion rates in headwater catchments affected by vegetation depletion (Chavasse, 1955; Holloway, 1959, 1966). A vigorous extermination campaign has achieved a degree of control over introduced animal numbers and recent research has questioned the role of opossums in causing mortality in rata-kamahi forests (Stewart and Veblen, 1981a, b; 1982a, b; Veblen and Stewart, 1982a). It is clear from Hawkes (1981) study and the distribution of precipitation across the Southern Alps (Griffiths and McSaveney, 1983a, b) that very high rates of erosion are related primarily to climatic and tectonic controls and the influence of man and introduced animals should be evaluated against an understanding of natural rates of erosion.

There are three major reasons for the paucity of quantitative data on rates and processes of soil erosion in the mountain lands. The principal difficulty lies with assessment techniques (see Chapter 2 for further discussion of assessment techniques). Much of the published work has relied on superficial reconnaissance surveys that use subjective ranking techniques to assess "severity", "seriousness" or "degree" of erosion (Gibbs *et al*, 1945; MWD, 1974, 1979; Cuff, 1974, 1977, 1981; Simpson, 1979). These assessments portray quantitatively the extent of erosion forms but this is not necessarily related to rates of soil erosion (Hawley, 1985). The historic photo-comparisons of Whitehouse (1978, 1982a, b) indicate difficulties in using the areal extent of erosion forms as an index of rates of erosion. Ranking techniques have been refined to provide estimates of point sources of sediment to streams (Cuff, 1974, 1977, 1981; MWD, 1977b; Simpson, 1979; Mosley, 1980; Whitehouse and McSaveney, 1980; McSaveney and Whitehouse, 1983).

Another major difficulty lies in distinguishing geological erosion from "accelerated" erosion. The Land Use Capability Survey Handbook (MWD, 1974) establishes classes of geological and "accelerated" erosion but gives no guidance on how the two might be distinguished. How much erosion and sediment transport is normal, and how much is accelerated by man's activities has yet to be quantitatively determined (O'Connor, 1976).

A further practical difficulty inherent in any study of erosion is temporal and spatial variability of erosion processes. The results of detailed process studies are difficult to extrapolate through time and space (Hayward, 1967, 1968, 1969; Boughton, 1967) while the results of any short term measurement programme are difficult to extrapolate through time.

As a result of these difficulties methods of assessment are being developed which use soil stratigraphy to assess erosion with respect to environment (geology, climate, topography, vegetation) and historical factors (tectonic history, climatic change in the Pleistocene and Aranuian, modification of vegetation cover). This approach is illustrated by the studies of Molloy (1964), Laffan and Cutler (1977), Archer (1978), Tonkin *et al* (1981), Harrison (1982), and Basher and Tonkin (1985). The incorporation of history is greatly facilitated by dating techniques such as radiocarbon dating, weathering rind thickness (Chinn, 1981), dendrochronology (Dunwiddie, 1979), and the comparison of historic and modern photographs (Whitehouse, 1978, 1982a, b). This soil stratigraphic approach assesses erosion in a medium term perspective (100-10,000 years).

Research is also tending to be based on whole catchment, multidisciplinary studies (e.g., Hayward, 1980; MWD, 1977a) with the recognition that the drainage basin is a convenient unit for geomorphic studies (Schumm, 1977; Gerrard 1981). Reconnaissance surveys do not allow the dynamics of soil erosion and sediment transport processes to be investigated at the drainage basin level, while localised, detailed process studies are difficult to extrapolate over entire drainage basins because of the inherent variability of erosion processes.

This brief review outlines the present understanding of erosion in the Southern Alps and the techniques that have been used to assess erosion. There is a lack of knowledge of rates and processes of erosion in areas west of the Main Divide. In view of the relationship between rainfall and sediment yield (Thompson and Adams, 1979; Griffiths, 1979b, 1981) and the distribution of precipitation across the Southern Alps (Griffiths and McSaveney, 1983b) this study was initiated to assess the effect of erosion on an essentially undisturbed (by fire, animal grazing, vegetation change) drainage basin within the zone of maximum precipitation in the central Southern Alps.

1.3 OBJECTIVES

This study had both general and specific objectives. In general terms studies in the extreme environment of Cropp River probably provide an upper limit to rates of erosion under natural vegetation in New Zealand's mountain lands. Soil studies provide information on soil development processes, and the resultant soil properties, characteristic of an area of extreme precipitation.

The specific aims of this study were:

1. To determine rates and processes of soil erosion in the drainage basin of Cropp River as a case study of erosional modification of mountainous schist terrain, subject to high rates of tectonic uplift and a superhumid climate;

2. To advance models of soil development and thereby elucidate those soil factors that:
   (a) contribute to vegetation and soil instability, and
   (b) can be reliably used in the recognition of chronological arrays of geomorphic surfaces;

3. To reconstruct the erosional history of Cropp River drainage basin.

CHAPTER 2 - REVIEW OF LITERATURE

2.1 INTRODUCTION

This is a selected review of the literature relevant to the general themes of this thesis i.e. dating techniques, soil stratigraphy, erosion assessment. Dating techniques are reviewed and their applications and limitations, within the scope of this study, discussed. The emphasis in this study has been directed toward using soils and soil stratigraphy to investigate erosion history. The use of soil stratigraphy requires a thorough understanding of soil development and the factors controlling soil distribution. Models of soil development and distribution are presented which provide a theoretical basis for interpreting observed soil patterns. To understand soil development, it is essential that features of the soil profile that are of sedimentary origin are distinguished from those of pedogenic origin. Techniques for evaluating the uniformity of parent material and the degree of soil profile development are outlined. Previous soil studies in Westland are reviewed since they provide an understanding of soil development and distribution in high rainfall environments. Erosion assessment techniques are described and terminology for the description of erosion forms introduced.

2.2 DATING TECHNIQUES

Dating techniques have been described as absolute, where they provide estimates of age in years, or relative, where age is given only relative to the age of some other feature or features (Ruhe, 1969a). The term "absolute" implies a precise knowledge of age and is not used here, since the techniques commonly described as absolute (radiometric dating and dendrochronology) are subject to errors, which make it difficult to ascribe calendar year ages to such dates. A more useful distinction can be made between numerical (equivalent to absolute) and relative dating methods. Numerical dating techniques can be further subdivided into direct and indirect methods. ^{14}C dating and dendrochronology are direct, numerical techniques where the results of measurement (numbers of tree rings, beta particle counts) are calibrated directly with time. Lichenometry, rock weathering rind dating and soil development are relative dating techniques, although data can be transformed to provide numerical age estimates by calibrating lichen diameter/weathering rind thickness/soil development against ages determined by direct methods. They are thus indirect, numerical techniques.

2.2.1 Dendrochronology

Dendrochronology is the science of dating annual growth layers of woody plants and the exploitation of associated environmental information. Dendrogeomorphology is the term that has been used for geomorphic applications of dendrochronology (Shroder, 1980).

Vegetation responds to geomorphic processes in many ways: trees or shrubs may become inclined; rootwood or stemwood may shear or be abraded; stemwood may be buried or rootwood exposed; vegetation may be buried or removed. Each of these events produces growth responses that can be dated. These include reaction wood growth, growth supression, growth release, ring termination and new callous growth, sprouting, miscellaneous structural and morphological changes in external or internal wood character, and initiation of plant succession (Shroder, 1980).

Four different sample types may be taken: cross-cut, longtitudinal-cut or wedge-cut sections, and cores (Lawrence, 1950; Burrows and Burrows, 1976; Schweingruber *et al*, 1978). Preparation of specimens involves mounting of cores or sections, surfacing with a razor or sandpaper, and, if necessary, staining to deepen latewood cell colour.

At least three types of tree ring dating are possible:

(a) Ring width counts and measurement (by microscope or various machines);

(b) Ring density measurement (using X-ray techniques);

(c) Event-response plotting (plots the internal morphology of samples to produce accurate individual dates of phenomena).

Adequate replication and cross dating are essential because of the range of event responses that can occur, and the possibility of missing or multiple rings in any year for a particular plant.

The following analytical methods may be applied to tree ring measurement:

(a) Isolated dates (easy to obtain and reliable with adequate replication);

(b) Long term chronologies (require greater attention to ring width and density, adequate cross dating and replication are essential);

(c) Spatial analysis (spatial distribution of event responses provides information on process mechanics).

These methods have been used to date such diverse events as shoreline warping, faulting, volcanism, stream erosion/ deposition, channel migration, slope denudation, and the activity of snow avalanches, glaciers, sand dunes and coastlines - Alestalo (1971) and Shroder (1980) provide bibliographies of the extensive literature. Dendrochronology has also been widely applied to paleoclimatic reconstructions and studies of forest population dynamics. In New Zealand dendrochronology is still in its infancy although its potential has been recognised (Dunwiddie, 1979; Norton, 1983b). Geomorphic applications of dendrochronology are limited to the studies of Mark *et al* (1964) in dating landslides, Druce (1966) and Topping (1971) in dating volcanic eruptions, Wardle (1973a) in dating glacial activity, and Grant (1981, 1985) and Smith and Lee (1984) in dating episodes of river terrace formation.

2.2.2 Radiocarbon dating

The method of radiocarbon dating has been widely used to date organic materials ranging in age from 200 to approximately 50,000 years before present (B.P.). With the technique of isotopic enrichment the dating range can now be extended to 75,000 years B.P. (Grootes, 1978). Applications of the method to earth science studies are numerous and no attempt has been made to review the extensive literature.

Radiocarbon (^{14}C) is continuously produced in the atmosphere where neutrons produced by cosmic rays react with nitrogen:

$$^{14}_{7}N + n \rightarrow {}^{14}_{6}C + p$$

Once formed the radioactive atoms are rapidly oxidised to CO_2 and introduced into the dynamic global carbon cycle, where an equilibrium concentration of ^{14}C is maintained by the balance between cosmic ray induced production and radioactive decay. When an organism dies its ^{14}C is no longer replenished. ^{14}C then decays to ^{14}N with the production of a beta (β) particle:

$$^{14}_{6}C \rightarrow {}^{14}_{7}N + \beta$$

The number of ^{14}C atoms remaining declines exponentially with time, with a half-life of 5730 $\pm$ 40 years. Thus a comparison of residual ^{14}C in the sample (A) with that for modern material (A_o) provides a measure of the time (t) elapsed since sample death according to the radioactive decay equation:

$$A = A_o.e^{-xt}$$

where x is the radioactive decay constant for ^{14}C ($\log_e 2$/half-life). The amount of ^{14}C in a sample is generally determined by measuring its beta-particle activity.

Assumptions inherent in ^{14}C dating are that:

- the biospheric carbon reservoir is in equilibrium;
- atmospheric ^{14}C level has remained constant;
- the decay rate of ^{14}C has been constant;
- no "young" or "old" carbon has been added to the sample since it was isolated from the global equilibrium;
- no istotopic fractionation has occurred to alter the standard ^{14}C:^{13}C:^{12}C ratios in the sample.

However, it is known that these assumptions are not completely valid and corrections are applied for some of the uncertainties inherent in ^{14}C date calculation. These include:

(a) Uncertainties in ^{14}C content at the time of death.

(i) Natural variation in atmospheric ^{14}C content (Secular effect).

Both long term (10,000 year) and short term (200 year) variations in ^{14}C content have been demonstrated from a comparison of the radiocarbon time scale and other dating methods (De Vries 1958, 1959; Yang and Fairhall, 1972; Damon *et al*, 1978; Stuiver, 1978; Suess, 1980). Attempts have been made to calibrate the last 8000 years of the radiocarbon timescale by dendrochronology (see Klein *et al*, 1982) but because of uncertainty in collating the various data there is, at present, no generally agreed calibration. This effect causes radiocarbon ages to be 800 years too young by 7000 years B.P., and to deviate by a maximum of 2000 years over the 9,000 to 32,000 years B.P. interval (Stuiver, 1978).

(ii) Artificially induced changes in atmospheric ^{14}C levels.

These include the Industrial or Suess effect (burning of fossil fuels with no residual ^{14}C has diluted the atmospheric ^{14}C concentration) and the Atom bomb effect (formation of large amounts of ^{14}C in the atmosphere by neutrons produced during atmospheric testing of atomic weapons). These effects are overcome by adjusting the contemporary activity of modern standards with reference to wood that is more than 100 years old. The error involved in cross calibration is incorporated into the standard error of dates.

(iii) Natural variations in ^{14}C content of the carbon reservoirs.

Isotopic fractionation. This occurs during the geochemical transfer of carbon in nature. The extent of enrichment on the $^{14}C/^{12}C$ ratio is approximately double that which may be directly measured for $^{13}C/^{12}C$. Enrichment data are expressed relative to a primary standard known as PDB (Cretaceous belemnites from the Peedee formation of South Carolina). ^{14}C activities are normalised to correspond to an enrichment value representing the mean isotopic composition of wood (-25‰ with respect to PDB). Reservoir Effect. It is assumed ^{14}C generated in the atmosphere is rapidly mixed with other carbon reservoirs to result in a uniform global specific activity. Most terrestrial life accurately reflects changes in atmospheric ^{14}C content but this equilibrium situation does not occur in marine carbon because of the enormous volume of ocean water and its slow circulation (Baxter and Walton, 1971). A significant amount of radioactive decay proceeds in the ocean and an "apparent" age is imparted to living organisms (the Reservoir Effect). For marine samples corrections are applied, according to latitude and ocean depth, from measurements on samples of known age.

(b) Uncertainties in the establishment of the half life of ^{14}C.

Accurate assessment of the half-life of ^{14}C is essential if radiocarbon ages are to be correlated with solar years. Libby (1955) obtained a value of 5568 ± 30 years ("Libby" half-life) from the mean of the three most reliable measurements then available. A value that is almost three percent higher (5730 ± 40 years) is now accepted as more accurate (Godwin, 1962; Olsson, 1968). By international convention laboratories continue to calculate and report "conventional" radiocarbon ages on the basis of the "Libby" half-life.

(c) Uncertainties in measuring sample ^{14}C activities.

The age of a sample is determined by the ratio of its activity to the activity of a contemporary standard. The activity of individual samples is expressed as the mean of a number of repeated measurements of particle activity. Uncertainties stem from the need to obtain a statistical mean value rather than an absolute measurement. The error term associated with dates includes the total uncertainty in measuring the activity of the sample, the contemporary standard, and the background. It is based solely on counting statistics and does not represent the true error limits of the age of the sample. Recent work (International Study Group, 1982) has indicated that quoted errors may be too low by a factor of 2 or 3 and urges caution in interpreting ^{14}C age differences of less than 200 years.

In view of these sources of uncertainty affecting radiocarbon dates,the term "conventional" radiocarbon age has been instigated to enable dates to be compared world-wide. The term "conventional" radiocarbon age implies:

- the use of the 5,568 year half-life;
- the assumption of constancy of ^{14}C atmospheric levels during the past;
- the use of oxalic acid (directly or indirectly) as a standard;
- isotopic fractionation normalisation of all sample activities to the base of $\delta^{13}C = -25‰$ relative to the $^{13}C/^{12}C$ ratio of PDB;
- the year 1950 is automatically the base year, with ages given in years B.P. (i.e. present is A.D. 1950).

In comparing two ^{14}C dates the actual value of the half-life is unimportant so long as the same value has been used to calculate both dates. When an accurate relationship with calendar dates is required, the most reliable value of the half-life is used and corrections for secular effects applied if these are known (McFadgen, 1982). Dates obtained in this study of Cropp River are discussed in terms of "conventional" radiocarbon age.

Radiocarbon dates are based on a best estimate of the ^{14}C content of the sample submitted to the laboratory. Additional and often unassessed sources of error in dating events by radiocarbon age determination can arise from:

(a) misassociation of sample and event (i.e. sample must be contemporaneous with the event being dated);

(b) sample contamination
- natural contamination by carbon of a different age to that of the sample;
- artificial contamination during collection, storage or subsequent packaging.

Care during field collection and sample preparation, and multiple sampling helps minimise problems of this nature. The magnitude of these problems depends largely on the type of material being dated and its age.

Materials suitable for radiocarbon dating include wood, charcoal, plant matter, peat, soil, organic mud (gyttja or dy), skin, hair, flesh, bones, skeletal carbonates and speleotherms. The reliability of radiocarbon dates varies according to the nature of the sample, its age and the environment from which it was collected. In the Cropp study only wood and peat have been used for dating and all samples were relatively young (<11,000 years B.P.).

Wood provides excellent material for dating as it does not readily exchange with contemporary carbon. Translocation of carbon compounds across tree rings (Long *et al*, 1976; Olsson, 1976) induces a maximum error of 2‰ (c. 15 years) for trees growing before the bomb era and is negligible for geochronological studies. Pretreatments to remove possible contaminants in woody material are not normally performed. Greater errors are likely from misassociation of sample and event. Twigs or young branches provide meaningful dates while heartwood from large trees, and relict timber (transported or *in situ)* may provide dates significantly older than the event (inbuilt age - McFadgen, 1982).

Peat is considered a reliable dating material by most laboratories, especially since it is normally formed *in situ*. However contamination by younger organic carbon compounds has been demonstrated for some peats (Grant-Taylor, 1972; Goh, 1978; Goh *et al*, 1978). Identification of the mobile contaminants and a routine procedure that gives reliable and consistent results is still the subject of debate (see Pullar, 1979; McGill, 1979; Grant-Taylor, 1979; Goh, 1979) and therefore most laboratories restrict pretreatment to an acid wash to remove carbonates (Grant-Taylor and Rafter, 1971). The magnitude of contamination error increases with sample age and degree of contamination (Jansen, 1984; Fig. 3). Published studies of peats from New Zealand (Goh, 1978; Goh *et al*, 1978) show increases in radiocarbon age, following pretreatment, ranging from about 500 years (for a sample dated at 9610 years B.P.) to 15,000 years (for a sample dated at 34,400 years B.P.).

2.2.3 Soils

Various soil properties change with time and indicate relative age. These include both morphological properties (e.g. sequence and thickness of horizons, texture or particle size distribution, B horizon colour) and chemical properties (e.g. pH, base saturation, oxalate- and dithionite-citrate extractable Fe and Al, variation in P fractions). Soil properties that are useful as indicators of relative age vary with climatic environment and are discussed by Evans (1982), Birkeland (1974, 1984b) and Richards *et al* (1985). Soils have often been used to support reconstructions based on geomorphology, stratigraphy, numerical and relative dating techniques (e.g. lichenometry, weathering rind development). The use of soils as a relative dating technique is part of the wider field of soil stratigraphy which is discussed more fully in the next section (2.3).

Soil development has been widely used as a dating technique and correlative tool in studies of Quaternary stratigraphy in North America. Numerous studies have been undertaken in the

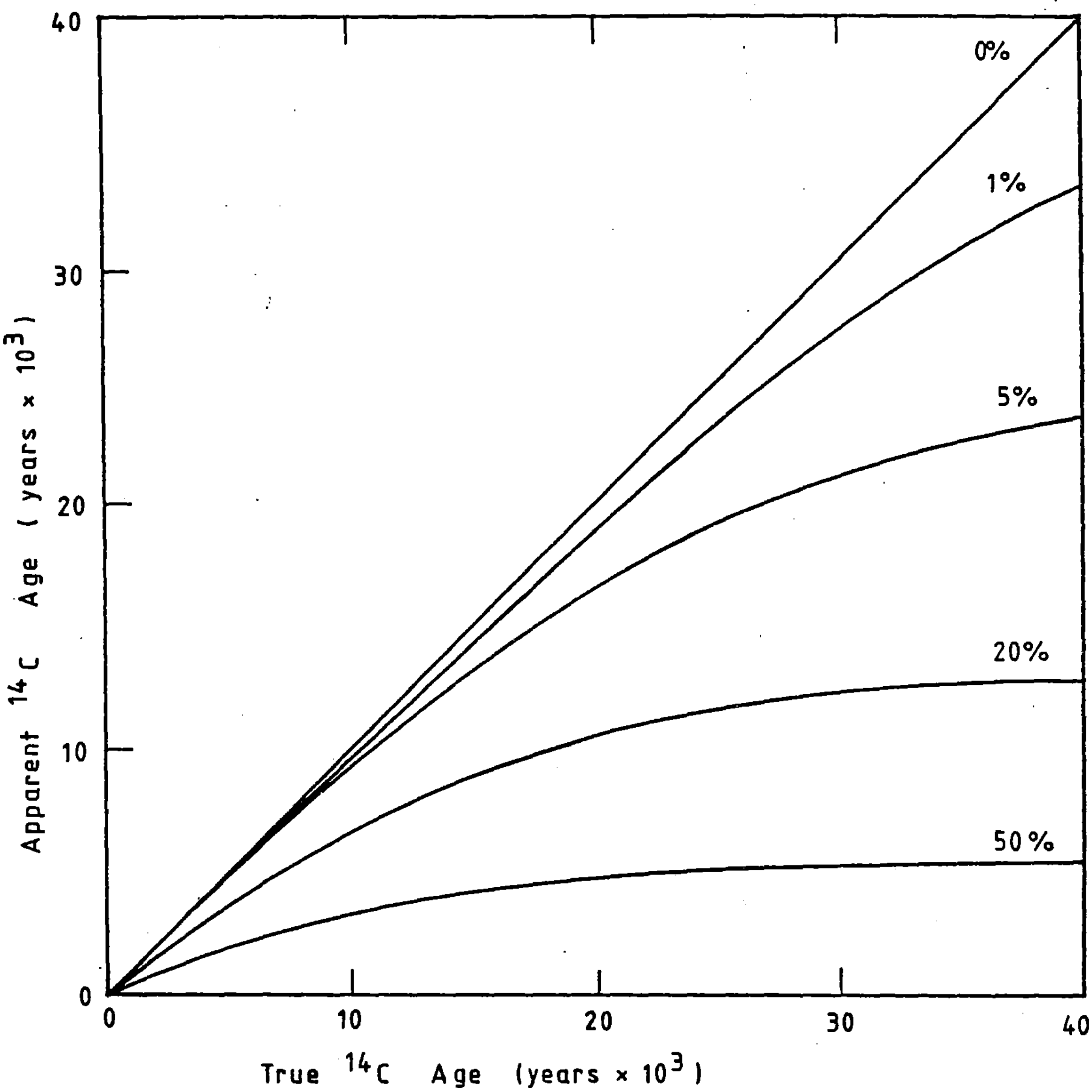


FIG. 3: Effect of contamination by modern carbon on ^{14}C dates (from Jansen 1984).

North American mountain ranges on moraine, outwash and rock glacier chronologies (e.g. Richmond, 1962; Birkeland, 1968, 1973; Mahaney, 1973, 1974, 1975, 1978; Burke and Birkeland, 1979; Birkeland *et al*, 1980). In the arid and semi-arid basin and range province of the south-western United States of America soil development on terrace and fan sequences was used in the correlation of Pleistocene and Holocene geomorphic surfaces over a wide geographical area (Gile and Hawley, 1968; Gile, 1975, 1977; Shlemon, 1978; Gile *et al*, 1981). Similar studies have been undertaken in the western USA deserts (Morrison, 1964, 1965), the southeastern USA coastal plain (Daniels *et al*, 1970; Daniels and Gamble, 1978) and in Oregon (Parsons *et al*, 1970; Parsons and Herriman, 1976).

Soils have also been used to differentiate glacial deposits in the Arctic (Birkeland, 1978; Bockheim, 1979) and Antarctic (Everett, 1971; Campbell and Claridge, 1975). In New Zealand soil properties have been used to investigate the age relationships of various landforms including terraces (Cox and Mead, 1963; Webb *et al*, 1981), moraines (Birkeland, 1981, 1982, 1984a; Gellatly, 1982 ; Somerville *et al*, 1982) and mountain hillslopes (Tonkin *et al*, 1981; Harrison, 1982).

2.2.4 Multiparameter techniques

The most reliable indications of relative age come from the use of a combination of relative dating techniques (Birkeland, 1973). The multiparameter approach reduces the possibility of mistakes in subdivision and correlation due to spatial (non-temporal) variation of a key parameter.

Many of the studies cited in section 2.2.3 are examples of the multiparameter approach. Most commonly a combination of numerical dating and stratigraphy in association with some or all of the following is used:

- rock weathering characteristics (rind thickness, ratios of fresh to weathered boulders, quartz vein height, corner angularity, surface pitting, surface oxidation and

many more);
- lichenometry;
- soil morphological and chemical properties.

For examples see Birkeland (1973, 1982), Burke and Birkeland (1979), and Gellatly (1982).

2.3 SOIL STRATIGRAPHY

Soil stratigraphy concerns the chronological ordering of pedological episodes (Finkl, 1980). It involves the study of soil development in relation to similarly aged erosional and depostional facets of the land surface, and the interpretation of stratigraphic layering and associated buried soils in the regolith i.e. soil stratigraphy involves study of, and determining relationships between, both surface and buried soils. Assessment of the stratigraphic significance of soils requires a throrough understanding of the processes responsible for soil development and the factors determining soil distribution. This involves distinguishing features of soil profiles that are of geologic origin from those that are of pedologic origin, and distinguishing soil patterns that are temporally determined from those that are spatially determined (Burns and Tonkin, 1982).

2.3.1 Models of soil development and distribution

Soils result from the interaction of geomorphologic and pedologic processes over time. The soil system, viewed within the landscape setting of a drainage basin, forms part of an open system, receiving and losing energy and matter at its boundaries (Jenny, 1980). Because the soil behaves as a complex open system it has commonly been modelled in terms of smaller subsystems of lesser complexity. The role and types of models that have been used in pedological studies are discussed in Dijkerman (1974), Huggett (1975), Smeck *et al* (1983) and Tonkin (1984). This section outlines conceptual models relevant to an understanding of soil development and distribution, within the scope of the Cropp study.

<u>2.3.1.1 Jenny's state factor model</u>

The soil system is a component of the ecosystem and responds to energy and mass fluxes at the ecosystem boundaries. The factors that control the state of the soil system are embodied in the state factor equation of Jenny (1941):

S (or s) = f(*cl*, *o*, *r*, *p*, *t*, ...)

where S denotes the soil, s any soil property, *cl* the regional climate, *o* the biotic factor, *r* the relief or topography, *p* the parent material, *t* the time factor, and the dots after *t* represent unspecified factors that may be important locally. As the factors are theoretically independent, a change in any one factor results in a change in the soil. In reality it is extremely difficult to find field sites in which the factors vary independently of each other. Often variation in one state factor causes variations in the others i.e. it is the interaction of the five soil forming factors that leads to definite conditions of soil formation. The state factors climate, organisms, topography and parent material are responsible for the spatial arrangement of soils of any given age and are often covariant. The state factor time dictates the temporal arrangement of soils in the landscape, and is the only truly independent variable in the state factor equation (Stephens, 1947; Chesworth, 1973a).

Jenny's concepts have been critically reviewed by a number of authors (Stephens, 1947; Major, 1951; Crocker, 1952; Stevens and Walker, 1970) and there have been attempts to reformulate and quantitatively solve the state factor equation (Jenny 1961, 1980; Yaalon, 1975; Bockheim, 1980). But the main difficulty is that although the state factor equation is expressed in a mathematical way it is essentially a verbal model (Dijkerman, 1974). However, as Birkeland (1974) notes: "although one may question how quantitative the use of factors is, it (the state factor equation) does provide a qualitative expression of the trends one can expect in an area". Such qualitative expressions of the variables of the state factor equation give rise to sequences rather than functions.

The study of soil sequences based on the soil forming factors has had a major impact on pedological research. Jenny's model stands as one of the main integrating concepts in pedology and has been important in ordering concepts of soil genesis. In particular the chronosequence and toposequence concepts have proved essential in understanding soil development and distribution in hilly landscapes.

2.3.1.2 The chronosequence concept

A chronosequence is a sequence of soils developed on similar parent materials and relief, under the influence of constant or ineffectively varying climate and biotic factors, whose differences can thus be ascribed to the lapse of differing increments of time since the initiation of soil formation (Stevens and Walker, 1970). The concept and its uses are reviewed in Jenny (1941, 1980), Rode (1961), Stevens and Walker (1970), Birkeland (1974), and Vreeken (1975a).

Stevens and Walker (1970) divided chronosequence studies into non-strict chronosequences (where constancy of the soil forming factors was not sufficiently established, or it was impossible to date ground surfaces accurately) and strict chronosequences (where constancy of the soil forming factors and ground surface dating was probably achieved). Strict chronosequences have also been called chronofunctions (Jenny, 1961; Bockheim, 1980). Strict chronosequences are rare in nature and therefore many reported studies are non-strict chronosequences.

In the majority of chronosequence studies soils of different age, from different locations, are arranged into a time sequence - the comparative geographical technique (Rode, 1961). It is assumed that these soils have, in the past, gone through stages characterised by the preceding members of the sequence. Vreeken (1975a) classifies the principal kinds of chronosequences in terms of the moments of inception and cessation of soil development (Fig. 4):

(a) Post-incisive - an array of soils that began forming at different moments in the past, that are either still exposed or were buried simultaneously;

(a) Principal kinds of chronosequences.

Chronosequence		Inception of soil development	Cessation of soil development	Soil history
POST-INCISIVE		Time-transgressive	Isochronous	Partial overlap
PRE-INCISIVE		Isochronous	Time-transgressive	Partial overlap
TIME TRANS-GRESSIVE	with partial historical overlap	Time-transgressive	Time-transgressive	Partial overlap
	without partial historical overlap	Time-transgressive	Time-transgressive	No overlap

(b) Schematic time diagrams for principal kinds of chronosequences. The bars represent time intervals of soil development.

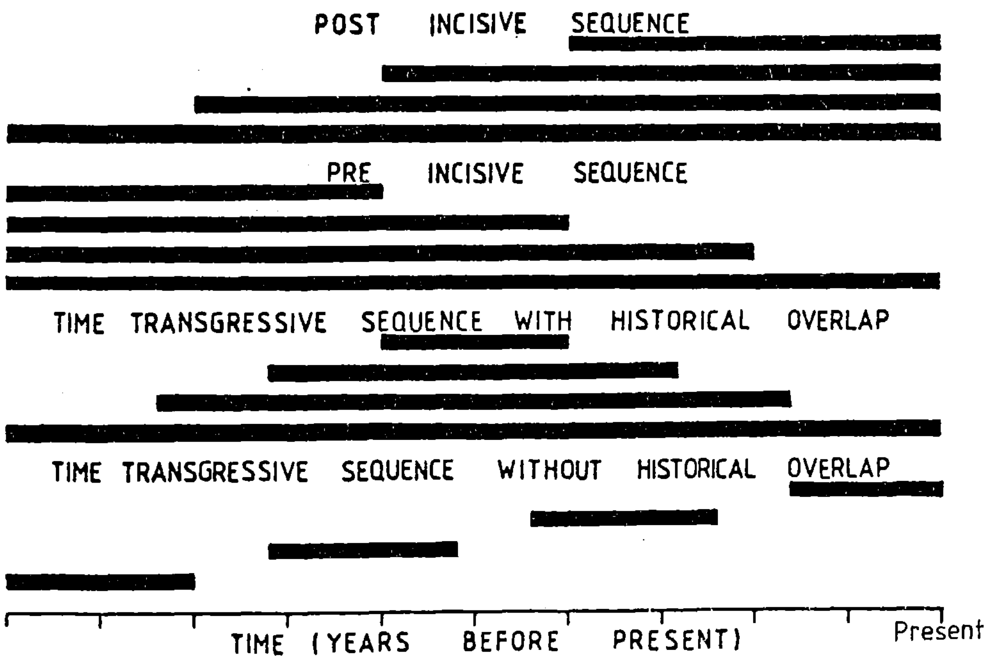


FIG. 4: Classification of chronosequences (from Vreeken, 1975a).

(b) Pre-incisive - an array of soils that began forming sim~~a~~ultaneously, but were buried at successively more recent moments, and may have one soil at the surface;

(c) Time-transgressive - arrays of soils that both began forming and were buried at different moments in time. Subdivisions with and without partial historical overlap are recognised.

Post-incisive sequences have commonly been used for inferences on pathways and rates of development of individual soils, on the assumption that all soils involved developed the same way and that macro-environmental conditions and local geomorphic regimes did not change significantly during the time interval encompassed by the sequence. Vreeken (1975a) acknowledges this as a problem, but recognises that the use of post-incisive chronosequences is necessitated by the nature of the stratigraphic and geomorphic record.

The study of soil chronosequences provides an empirical means of dating land surfaces of unknown age (2.2.3). The chronosequence concept is used to establish an *a priori* model of soil development with time, that then provides the basis for interpretation of surfaces that are not directly dateable. In applying soils to dating surfaces, comparability of sites and parent materials (for both reference sequence and undated surfaces) is crucial (Vreeken, 1984).

2.3.1.3 The toposequence concept

A toposequence is a sequence of related soils that differ, one from another, primarily as a result of the effect of topography on soil development. Soil properties change with variation in aspect, slope contour and profile. The concept of the soil catena provides a model of soil development on hillslopes, integrating geomorphological processes of erosion, transport and deposition, hydrological processes of overland flow and throughflow, with pedological processes (Tonkin, 1984). The original concept of the soil catena (Milne, 1935) was a sequence of spatially linked soils, extending from hill summit to valley floor, that was repeated across the contours of the landscape. The concept was extended to incorporate the properties of the three-dimensional soil landscape by Huggett (1975) who proposed the valley basin as the fundamental soil landscape unit for hilly landscapes. The valley basin is presented as a conceptual, homomorphic, structural model, the boundaries of which delimit a three-dimensional open system. This system forms part of a more extensive drainage network and possesses functional unity but not necessarily homogeneity of soil properties. The soil pattern within valley basins can be determined by analysis of the array of two dimensional soil catenas approximating the hydrological flow lines from watershed to thalweg (Tonkin *et al*, 1977). Soil profiles are differentiated vertically and valley basins and catenas horizontally. Catenas consist of:

(a) an (upper) eluvial zone of net solute and/or sediment loss;
(b) a (middle) transluvial zone of predominantly lateral solute and/or sediment transfer;
(c) a (lower) illuvial zone of partial solute and/or sediment accumulation.

The real significance of catenas lies in the recognition of the processes involved in soil differentiation on slopes, and of the interaction of pedological, geomorphological and hydrological processes. Of all the processes that relate soils to landscape geometry, the movement of water from slope summit to valley thalweg by various pathways is the most important. This conceptual model applies to both soil morphological properties (Tonkin *et al*, 1977) and chemical properties (Young *et al*, 1977).

Catenas also have temporal significance since a certain amount of time is required for catenary relationships to develop. As soils change with time so do landforms, and for catenary sequences to exist the relationships need to remain stable. The catena is a dynamic feature that changes with time and can be seen as an essential part of the processes of erosion and deposition (Gerrard, 1981).

Most catenas are open systems but they can also function as closed systems when runoff is into enclosed surface depressions (Walker and Ruhe, 1968; Malo *et al*, 1974). The closed system can be used to assess the relative transfers within component soil catenas, while comparative studies between open and closed systems provide a means of determining the relative magnitude of material losses (Ruhe and Walker, 1968; Walker and Ruhe, 1968).

Watson (1965), Ollier (1976) and Gerrard (1981) have reviewed the literature describing soil catenas formed in different climatic environments. Ollier (1976) draws a distinction between extreme and non-extreme situations. Distinctive soil-slope relationships occur in the extreme situations dominated by frigid (Tedrow, 1966) or arid (Dan *et al*, 1968) conditions. Under non-extreme conditions the processes of slope erosion, deposition and pedogenesis are interwoven. The universality of the soil catena model is demonstrated by studies in climatic environments from Arctic through temperate to tropical (Ollier, 1976; Gerrard, 1981).

While the catena concept is important in studying soil development on hillslopes, the considerable variety of slope form often makes comparison between soils and slopes extremely difficult. The advent of statistical techniques and the use of computers has enabled soil-slope relationships to be treated mathematically (Walker *et al*, 1968a, b; Furley, 1968, 1971, 1974a, b; Whitfield and Furley 1971; Vreeken, 1973). Statistically significant linear, curvilinear and polynomial relationships have been obtained for several soil properties but the solutions are of specific rather than general validity (Yaalon, 1975). The use of statistical techniques means that much greater attention must be given to sampling schemes since soils are extremely variable bodies. These studies show that steepness of slope affects soil properties (through its control on hillslope hydrology and erosion rates) and there is a fundamental difference between upper and lower slope positions (as a result of changing hydrological relationships and the balance between erosion and deposition). Slope position appears to be more important in determining soil properties than slope angle (Gerrard, 1981).

The direction faced by a slope (i.e. aspect) also has an important influence on soil properties. Aspect influences soil formation primarily through its effect on microclimate and vegetation. In theory the rate of soil development is a function of energy available to do work on the system, and the efficiency with which soluble reactant products are leached from the soil system (Yaalon, 1960). Relative rates of soil development with respect to aspect are therefore a function of soil water and soil temperature regimes. Where available water is non-limiting and temperature regimes are optimal, soil development should be faster on sunny aspects than shady (Losche *et al*, 1970). Conversely if available soil water is limiting soil development may proceed at a faster rate on shady aspects (Cuff, 1973; Webb, 1976; McIntosh *et al*, 1981). Many of the studies of soil development in relation to aspect appear to give conflicting results, but these conflicts can be resolved where more detailed information is provided on soil moisture and temperature regimes, and comparability of parent materials (Tonkin, 1984). Reviews of studies of soil development in relation to aspect may be found in Ross (1971), Cuff (1973), Webb (1976) and Tonkin (1984).

2.3.2 Terminology

Soil stratigraphy investigates the chronological ordering of pedological episodes expressed in surficial and buried soils. Because the stratigraphic interpretation of soils differs from conventional pedologic description, special terms have been invoked to describe the stratigraphic features of soil profiles and soil mantles.

2.3.2.1 Paleosols

A paleosol is a soil which has formed in a landscape of the past i.e. the former soil forming processes were either altered as a result of a change in the external environmental conditions or were interrupted as a result of burial (Yaalon, 1971). The stratigraphic occurrences of paleosols are described by the following terms (Morrison 1967, 1978):

(a) relict paleosols - have remained exposed at the surface since soil formation began;

(b) buried paleosols may be

(i) simple - where only a single buried soil is present;

(ii) compound - where 2 or more buried soils are present that do not, or only slightly overlap, in vertical succession;

(iii) composite - where 2 or more buried soils are present that overlap one another to the extent that the original horizons of the individual soils are difficult to distinguish;

(iv) subdivided - where a composite paleosol is traceable laterally to a compound paleosol;

(c) exhumed paleosols - a paleosol or remnant of a paleosol formed, buried, then subsequently exhumed by erosion.

These groups are members of a continuum and any stratigraphic succession may contain one or more types.

A number of authors have criticised the use of the term 'paleosol' because the meaning of 'landscape of the past' has never been well defined (Ruellan, 1971; Bos and Sevink, 1975; Catt, 1979; Kemp, 1985). Use of the term in relation to surface soils depends on recognition of profile features that cannot have developed under present-day conditions. Criteria for distinguishing 'modern' soils from 'relict' soils have never been established. Kemp points out that the term 'paleosol' qualified by an adjective (buried, relict or exhumed) implies no more than the term 'soil' qualified by the same adjective - and suggests use of the term 'paleosol' be discontinued. Thus 'paleosol' is not used in describing soils in Cropp basin, but the terms for describing the stratigraphic occurrence of buried soils are used (i.e. simple, compound, composite).

2.3.2.2 Soil stratigraphic units

Soil mantles used for correlation purposes are called soil stratigraphic (pedostratigraphic) units. These are soils with physical features and stratigraphic relations that permit their consistent recognition and mapping as stratigraphic units (American Commission on Stratigraphic Nomenclature, 1961; Brewer *et al*, 1970; Gage, 1977; Butler, 1982; North American Commission on Stratigraphic Nomenclature, 1983; Walker *et al*, 1984). Soil stratigraphic units may parallel or transgress time horizons. They may exhibit spatial variation in soil morphology due to variation in environmental conditions (e.g. parent material or climate may differ), but stratigraphic relationships must be consistent.

Several terms for soil stratigraphic units have been proposed including:

(a) the geomorphic surface (Ruhe, 1956, 1969a; Daniels *et al*, 1971; Gile *et al*, 1981);
(b) the ground surface (Butler, 1959);
(c) the geosol (Morrison, 1967, 1978; North American Commission on Stratigraphic Nomenclature, 1983);
(d) the pedoderm (Brewer *et al*, 1970; Butler, 1982; Walker *et al*, 1984; Beckmann, 1984).

Each of these terms describes a similar (but not identical) concept.

The geomorphic surface concept is suitable for detailed drainage basin studies where emphasis is on local chronology rather than regional stratigraphy, which would fulfill the requirements for formally recognising pedoderms or geosols. A geomorphic surface is a portion of the land surface, comprising both erosional and depositional elements, that is specifically defined in space and time (Ruhe, 1969a) i.e. it has definite geographic boundaries (is mappable) and is formed by one or more agencies over a defined time period. The term has been used to refer to a two dimensional surface (Ruhe, 1969a; Daniels *et al*, 1971; Daniels and Gamble, 1978) or a three dimensional weathering zone (Gile *et al*, 1981). Daniels and Gamble (1978) argue that although weathering of the underlying sediments can be part of the site history, the weathering zone is not the geomorphic surface. Geomorphic surfaces may occur at the present land surface or they may be buried. Transformation of geomorphic surfaces by erosion and deposition results in the formation of new geomorphic surfaces. Intergrade elements present during the transformation of geomorphic surfaces by erosion and deposition are described as degradational or aggradational phases of the present geomorphic surface (Tonkin *et al*, 1981).

2.3.3 Principles of soil stratigraphy

The relationship between soils and geomorphic surfaces and buried soils and buried geomorphic surfaces is the basis of soil stratigraphy (Morrison, 1978). Soil stratigraphy applies to soil mantles, not to soil profiles. Soil mantles, although extensive, are discontinuous and it is the nature of their discontinuities and contacts which is important (Butler, 1967). The principles of soil stratigraphy are discussed by Butler (1959), Richmond (1962), Morrison (1967, 1978), Ruhe (1969a, b), Daniels *et al* (1971), Finkl (1980), Tonkin *et al* (1981) and Vreeken (1984).

The key points are:

(a) Stratigraphic relationships between geomorphic surfaces and buried geomorphic surfaces may be resolved according to:

(i) the law of superposition i.e. younger surfaces occur on top of older surfaces providing they have not been overturned;

(ii) the law of cross-cutting relationships i.e. in a cross-cutting relationship, the surface that is cut is older than the surface that cuts across it;

(iii) the principles of ascendancy and descendency i.e. a surface may be the same age or younger than a higher surface to which it ascends; a surface is the same age as the alluvial fill to which it descends;

(b) Buried geomorphic surfaces are identified by the presence of buried soils. Dating of materials associated with buried soils can provide a minimum age for sediments that underlie a buried geomorphic surface and a maximum age for sediments that overlie the surface;

(c) The age of a geomorphic surface, its associated soils and surface weathering of exposed clasts are the same, so long as there is minimal addition to, or removal from, the surface;

(d) Soils associated with erosional and depositional elements of a geomorphic surface may differ, because of contrasts in regolith. These contrasts may be further accentuated by soil catenary relationships i.e. a soil stratigraphic unit may consist of several pedologic units but its stratigraphic relationships are constant;

(g) Maps showing the distribution and relative ages of geomorphic surfaces depict the mappable erosion and deposition history of a drainage basin. Areas of present surface instability can be identified because they lack a diagnostic soil, or other surficial weathering features of a geomorphic surface.

An important principle in soil stratigraphy is that for every depositional surface, indicated either by the soils or buried soils, there was a coeval erosional surface. Often this erosional surface is modified or destroyed during subsequent episodes of erosion. Therefore the best record of erosion history is to be found in the stratigraphy of the depositional components of the landscape, such as alluvial fans and valley fills. Buried soils are indicative of slope stability between periods of instability. The radiocarbon dating of organic plant remains, commonly associated with buried soils, can provide a chronology for episodes of erosion and deposition.

Soil stratigraphic principles have been most widely applied to the study of landscapes of considerable antiquity (>10,000 to

2 m years) with a relatively slow tempo of landscape evolution and infrequent episodic erosion (2.2.3). Soil stratigraphy can also be applied to the study of young landscapes with rapid rates of landform evolution and frequent episodic, or continuous, erosion. In such situations there are practical problems of mappability and temporal persistence of geomorphic surfaces, especially for erosional components of geomorphic surfaces. Despite these practical difficulties the concepts of soil stratigraphy are appropriate for investigating landscape history.

2.3.4 Quantitative techniques for evaluating soil profile development

The key to identifying chronological sequences of geomorphic surfaces is the recognition and characterisation of their associated soils. For surface soils (with which the Cropp study is primarily concerned) it is necessary to evaluate the degree of soil profile development of individual soils and to make comparisons between soils.

The first requirement is to establish the initial state of the soil system including both the uniformity of parent material within a soil, and the uniformity of parent materials between different soils. An important characteristic of the soil parent material is that its inherent variability should not be significant compared to changes brought about by pedogenesis. Identification of lithologic discontinuities and parent material variability are an essential prerequisite in differentiating soil properties that are pedogenic from those that are of geologic origin. Both field (stratigraphic/ morphologic) and laboratory (physicochemical) criteria may be used to assess parent material uniformity. Field observations commonly indicate major discontinuities in parent materials, particularly those relating to changes of permeability or pore size distribution (Wang and Arnold, 1973). These are often observed as changes in gravel content, fine earth texture, size and type of structure, consistence, and horizon boundaries. Physicochemical criteria used to assess parent material uniformity include the particle-size distribution of resistant minerals, or of the non-clay fraction; the ratio of two resistant minerals in any one size fraction; and the depth distribution of resistant minerals (Barshad, 1964; Evans, 1978). The main consideration for physicochemical criteria is that the substance(s) chosen has remained inert and immobile during pedogenesis. The nature of the parent material itself and the operative soil processes will dictate the most useful of these methods. Estimates of uniformity should be based on as many lines of evidence as possible (Evans, 1978).

The extent and manner in which the parent material is altered during pedogenesis can be determined by a variety of techniques. Numerical methods of rating soil morphological properties (soil development indices) are described by Bilzi and Ciolkosz (1977), Meixner and Singer (1981), Harden (1982), Harden and Taylor (1983) and Birkeland (1984a, b). Soil development indices convert descriptive field properties into numerical data to portray total profile change with time. These indices provide a qualitative measure of pedogenesis and are of value where quantitative data on soil chemical properties are not available. To date they have only been applied to the description of temporal soil variability. Their applicability to evaluating integrated spatial and temporal variability remains to be tested.

Numerical methods of rating chemical properties of soils have been widely used to evaluate profile development. Walker and Green (1976) propose an index of profile anisotropy (IPA) that compares the deviation of a sampled horizon (D) from the overall weighted mean value (M) of a particular soil property:

$$IPA = D \times 100/M$$

Birkeland (1984a, b) modifys this index by comparing the deviation of a sampled horizon from the value in the parent material:

$$mIPA = D/PM$$

Values are calculated for each horizon, then multiplied by horizon thickness and summed to give a profile value. Values are then normalised by dividing by profile depth. Because anisotropy generally increases with time, so should the index. However difficulties arise where anisotropy decreases with time, e.g. when the soil is made up of initially anisotropic material, or in very old soils where values of many chemical properties become isotropic as a result of extreme leaching and weathering. Furthermore these indices do not take account of bulk density changes during soil development or dilution by organic matter in upper horizons.

More quantitative estimates of nett change during soil development are made by assuming that certain minerals (referred to as internal standards) remain unaltered and immobile during soil development (Barshad, 1964; Brewer, 1964; Evans, 1978). Change can then be assessed relative to the internal standard. While no mineral is completely stable under all conditions, the aim is to choose the most stable and immobile constituent to minimise the error in evaluating profile development. Depth distributions of internal standards can also be used to evaluate parent material uniformity. A more detailed discussion of the application of internal standards to evaluation of soil development is presented in Chapter 6.

2.3.5 Applications of soil stratigraphy

Soil stratigraphy has been used in subdividing and correlating local and regional Quaternary sedimentary sequences, aiding paleoclimatic reconstructions and studying the evolution of landscapes (Gerrard, 1981).

Butler (1959, 1982) uses landscape periodicity as a framework for soil studies. He proposes the idea of the soil cycle (K-cycle) comprising the alternation of instability (when older surfaces are destroyed or buried) and stability (when soil development proceeds on the new surfaces). The evidence for K cycles are soil layers or "groundsurfaces" (all those erosional and depositional surfaces and layers which have developed in a landscape during one interval of time and upon which a unit mantle of soils has developed). Periodic landscape activity develops characteristic zones of ground-surface relationships (Fig. 5) called the persistent, accreting, alternating and sloughing zones. The K-cycle model conceptually describes the spatial relationships of ground surfaces and establishes criteria for recognising chronological sequences of erosional and depositional events. It provides a powerful model for the study of terrestrial erosion-deposition systems. The K-cycle model has been widely applied in Australia to investigate phases of slope instability, alluviation and landscape development (Butler, 1958, 1959, 1967, 1982; van Dijk, 1959; Walker, 1962a, b; Churchward, 1963; Beattie, 1972; Costin and Polach, 1973a, b; Walker and Green, 1976; Walker and Coventry, 1976). As yet, no regional chronology has been derived from these studies, although it is recognised that many of the cycles are probably climatically controlled. The K-cycle model has also been used to investigate geomorphic history of areas in New Zealand (Laffan and Culter, 1977; Archer, 1978), Spain (Mucher *et al*, 1972; Pereira *et al*, 1978) and Luxembourg (Kwaad and Mücher, 1977).

Soil stratigraphic principles have been extensively used in studies of loess stratigraphy throughout the world (see Pye, 1984). In the loess regions of Central Europe (particularly Czechoslovakia and Austria) loess sequences and intercalated buried soils, on river terraces close to the former fronts of the Scandanavian and Alpine ice sheets, record a sequence of 17 major glacial-interglacial shifts in the past 1.7 million yrs (Kukla, 1975; Fink and Kukla, 1977). Paleosols of well known soil stratigraphic horizons have also been used to investigate the complex history of loess deposition in the mid continental region of the USA (Ruhe 1969a, b; Ruhe *et al*, 1971; Vreeken, 1975b; Morrison, 1978; Follmer, 1978, 1982, 1983).

Relict and modern soils have been used to provide relative ages of surficial rock units and as a correlative tool in

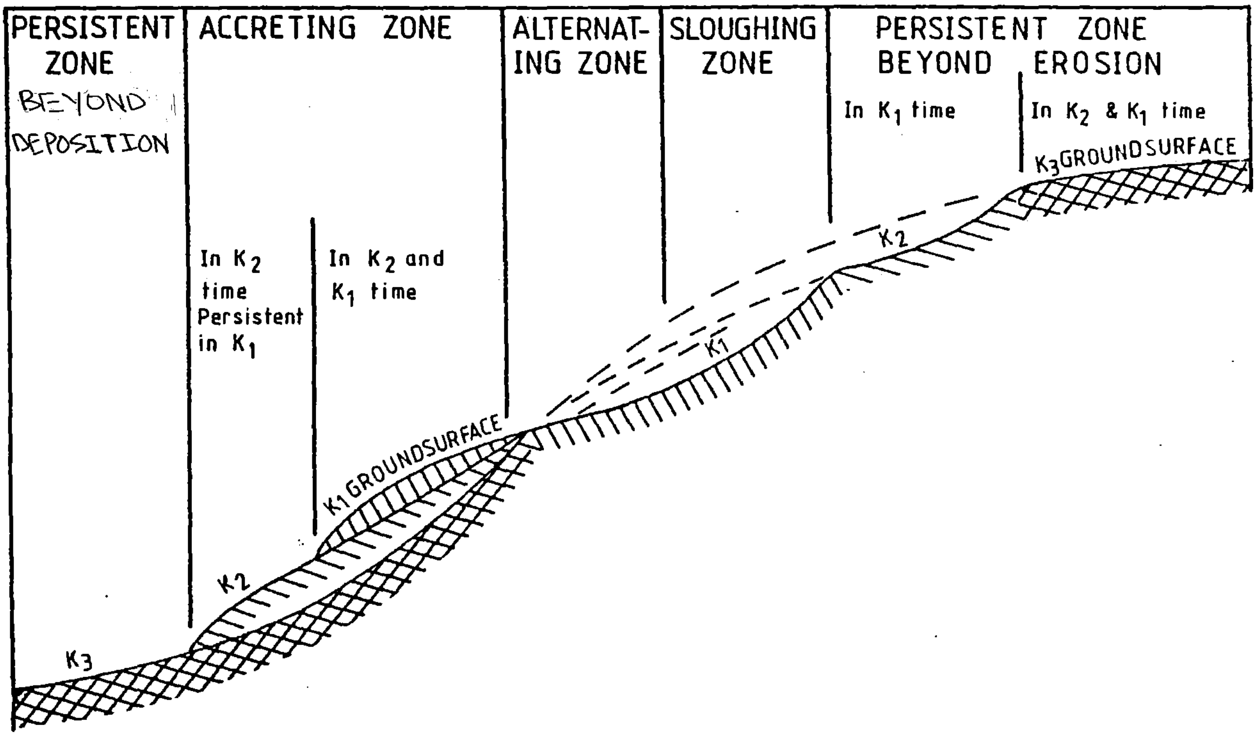


FIG. 5: The K-cycle model-diagrammatic illustration of K_1, K_2, K_3 ground surfaces on a hillslope, showing the relationship of groundsurface zones (from Butler, 1959).

studies of Quaternary moraine, rock glacier, terrace and fan chronologies (2.2.3).

A further application of soil stratigraphy is in the study of neotectonics and flood frequency. Soil stratigraphy has been used to evaluate the time of last movement of faults and to investigate the history of fault activity (Shlemon, 1975; Kirkham, 1977; Machette, 1978; Douglas, 1980; Swan *et al*, 1980). One of the greatest difficulties in flood frequency analysis is the estimation of the risk of rare, high magnitude floods. Stratigraphic, morphologic and pedologic evidence can be used to provide data on pre-historic large floods and to improve estimates of recurrence intervals for rare catastrophic floods (Costa 1974, 1978a, b).

Applications of soil stratigraphy in New Zealand include studies of tephrochronology, loess stratigraphy and a few detailed studies aimed at reconstructing geomorphic history. Extensive tephra deposits and buried soils have been utilised to analyse the history of volcanic activity and landscape evolution during the Quaternary (Pullar, 1967, 1973; Kennedy *et al*, 1978; Grant, 1981). The stratigraphic features of New Zealand's loess deposits have been studied and correlations with glacial advances and retreats suggested (e.g. Bruce, 1973a, b; Cowie and Milne, 1973; Griffiths, 1973; Ives, 1973; Runge *et al*, 1973; Milne, 1974; Tonkin *et al*, 1974). Problems in identifying buried soils, lack of significant direct numerical dates, and correlation of regional sequences have made development of a New Zealand-wide chronology difficult (Leamy *et al*, 1973). The principles of soil stratigraphy have been applied to the reconstruction of geomorphic history of the Otago Peninsula (Leslie, 1973a, b), Wither Hills (Laffan and Cutler, 1977), Ruahine Range (Hubbard and Neall, 1980) and the South Island mountain lands (Molloy, 1962, 1964; Ives, 1970; Harvey, 1974; Archer, 1978; Tonkin *et al*, 1981; Harrison, 1982; Basher and Tonkin, 1985). Evidence of soil periodicity relating to Quaternary tectonism, climatic and vegetation change is extremely widespread.

2.4 SOIL DEVELOPMENT AND DISTRIBUTION IN WESTLAND

Previous studies of soil development and distribution in Westland provide an understanding of the temporal and spatial factors that control the development of characteristic soil patterns in high rainfall environments. They provide a basis for interpreting soil patterns in Cropp basin.

Gibbs, Mercer and Collie (1950) describe the broad pattern of soil distribution in Westland. More detailed surveys have been undertaken by Mew *et al* (1975) in the Inangahua Depression, Mew (1980a, b) in the Greymouth-Hokitika region, Laffan (1980) in the Charleston-Punakaikai region, and Heine and Mew (1981) in the Mokihinui-Orikaka region. The broad findings from these surveys are summarised by Mew and Leamy (1977). Seven main groups of soils are recognised -recent soils on river flats, yellow brown sands on coastal dunes, yellow-brown earths on terrain ranging from low terraces to steep hillslopes, gley soils where watertables are permanently high, gley podzols on intermediate and high terraces, podzols on a variety of terrain, and organic soils in poorly drained areas. The chief pedological trends noted are:

(a) increasing incidence of gley soils on low glacial outwash terraces with increasing rainfall;
(b) an increase in gleying in hill and steepland soils with higher rainfall;
(c) an increasing tendency for soil and geological instability with increasing angle and length of slope on certain rock types, possibly coupled with higher rainfall and changes in land use;
(d) contrast in types and down-profile movement of organic matter associated with different forest types, mainly beech-podocarp and podocarp-hardwood forests.

These broad surveys provide a general account of the occurrence and properties of the major soil groups in Westland.

The development of soils in Westland has been examined in several chronosequence studies. The areas studied have rainfalls ranging from 1900 mm a^{-1} to 6500 mm a^{-1}. Successive stages of soil development are inferred from soil morphology, chemistry and mineralogy. Comparison of the studies indicates the effect of increasing rainfall on soil development. The observed sequences are characteristic of sites that are freely draining or, for the gley podzols, are in an equivalent topographic position. All soils are formed on flat or gently sloping topography.

The general characteristics of soil sequences on two terrace systems in the Inangahua and Grey valleys are described by Ross *et al* (1977). The soils are formed from post-glacial and glacial outwash gravels, with loess on the intermediate and high terraces, under a rainfall of 1900 mm a^{-1}. Recent soils are formed on the floodplains (Hokitika series from Recent alluvium), immature yellow-brown earths on low terraces (Ikamatua series from Aranuian alluvium), mature yellow-brown earths on higher low terraces (Ahaura series from Otiran alluvium), gley podzols on intermediate and high terraces (Okarito series from Waimean and Waimaungan alluvium). Accompanying the change in soil morphology is a general decline in soil fertility from lower to higher terraces.

More detailed studies of soil development on one terrace sequence in the Inangahua valley near Reefton are described by Tan (1971) and Campbell (1974, 1975). The terraces range in estimated age from 1000 years to greater than 130,000 years, although there are no direct dates to support these ages. The sequence includes recent soils (1000 years, Hokitika soils) yellow-brown earths, podzolised and gleyed yellow-brown earths and podzols (14,000 to 18,000 years, Ikamatua and Ahaura soils) and gley podzols (70,000 to 130,000 years?, Kumara and Okarito soils. These Kumara soils would now be mapped as Okarito soils -Ross *et al*, 1977). Variations in pH, organic matter, particle size, cation exchange properties, total Mg, Al, Si, K, Fe and Ti, poorly-ordered and organic-complexed forms of Fe and Al, and mineralogy caused by increasing duration of weathering and by short-term, short-range variations in the intensity of the biotic factor are described. Campbell (1975) concludes that the younger soils represent dynamic systems in which alternative weathering cycles can replace each other as the growth, death and eventual disappearance of individual red beech (*Nothofagus fusca*) trees caused localised fluctuations in pH. These processes eventually led to the formation of gley podzols, now found on the two oldest surfaces, with podzolisation preceding gleying. Campbell (1975) considers that the presence of loess of low permeability influences soil development, but it is not a necessary prerequisite for the formation of gley podzols. Gley podzols are characteristic of terraces believed to be older than the youngest advances of the Otira Glaciation (i.e. at least 22,000 years and possibly as old as 70,000 years).

A chronosequence of soils on terraces and moraines associated with the retreating Franz Josef Glacier is reported in Stevens (1963, 1968) and Mokma *et al* (1973). The surfaces range in age from 0 to an estimated 22,000 years and comprise coarse textured alluvium and till. Rainfall was estimated to range from 3800 to 5800 mm, although more recent study (Griffiths and McSaveney, 1983b) shows rainfall is more variable. This sequence shows the progressive development of recent soils (<100 years), yellow-brown earths (250 to 1000 years) and gley podzols (>5000 years). Some of the younger soils were also gleyed, possibly as a result of the low permeability of the till from which they were derived. Variation in pH, organic matter, particle size, total Mg, Ca, K and S, poorly-ordered forms of Fe and Al, exchange chemistry, and forms and amounts of P are outlined. Soils were first podzolised and then gleyed in some horizons. A gley podzol was formed within 5000 years. Trends at Franz Josef were similar to those obtained at Reefton, but the rate of soil development was faster and there were more surfaces to characterise the early stages of soil development at Franz Josef. The end members of both sequences were very similar i.e. Okarito gley podzol.

Soil development on river terrace sequences under a 6500 mm annual rainfall has been examined at the Arawata River (Smith and Lee, 1984) and Wanganui River (Sowden, 1986). Soils of both sequences are formed from sandy alluvium. At the Arawata River there are recent soils on the low terrace (150 years), podzolised yellow-brown earths on the intermediate terrace (450 years) and podzols on the high terrace (1400 years). Gleying is evident on soils on the intermediate and high terraces, and the podzol on the high terrace has an iron pan. At the Wanganui River there is a sequence of eight post-glacial terraces that has a similar sequence of soils to the Arawata. The sequence includes recent soils (<250 years), yellow-brown earths (600 to 1000 years), podzolised yellow-brown earths (1500 years), and gley podzols (2500 to 8000 years). Both sequences show rapid organic matter build up and early podzolisation. The rapid rate of soil development results from the combination of high rainfall and fine-textured parent material.

These soil sequence~~s~~ studies show a consistent sequence of soil development. As rainfall increases, the rate of soil development increases and there is greater influence of early gleying. The major kinds of physical and chemical changes with increasing soil development are:

(a) A decrease in the proportion of stones and gravels, and an increase in silt and clay in surface soil horizons. In the most acid soils the proportion of clay decreases due to clay destruction;

(b) Progressive acidification of surface soil horizons;

(c) Rapid initial increases in C, N, and C.E.C., correlated with the accumulation of organic matter. This was followed by a decrease in these parameters. Base cations showed a similar trend;

(d) An initial increase in poorly-ordered forms of Fe and Al in surface horizons. As surface horizons are acidified poorly-ordered forms of Fe and Al are dissolved and reprecipitated in the B horizon. The zone of maximum accumulation moves down the profile with time;

(e) Loss of total P and the transformation of primary forms (e.g. apatite) to secondary organic and inorganic forms. A model of phosphorus transformations as an index of soil development has been developed from a number of studies, including the Franz Josef chronosequence (Fig. 6 - Walker and Syers, 1976);

(f) Decrease in amounts of total Ca, Mg, K, Al and Fe.

The properties and genesis of eight of the most widespread soil series of the West Coast wetlands on low, intermediate, and high terraces, are described in a series of recent papers (Mew and Lee, 1981; Thomas and Lee, 1983; Mew *et al*, 1983; Barratt, 1983; Farmer *et al*, 1984; Jackson, 1984). All the soils examined are very strongly developed (podzols, gley podzols and gley soils) and many of their properties result from extreme wetness over long periods of time. They also exhibit characteristics attributable to podzolisation. Farmer *et al* (1984) suggest that podzolisation progressively forms an iron podzol, a humus iron podzol and a peaty podzol with a thin iron pan at the interface of B_h and B_s horizons. Permanent waterlogging above the iron pan then leads to local penetration of the pan by acidic reducing waters, and formation of discontinuous iron pans at greater depths in the older gley podzols. Jackson (1984) notes that waterlogging in these soils may be due to the properties of the fine textured upper horizons (low permeability, lack of large pores) rather than the presence of an iron pan.

Two detailed studies of soil distribution in Westland show that topography is also a major influence on soil development. Soil variability within the Blackball hill soils in some small catchments near Reefton (rainfall 1900 mm a^{-1}) is described by McKie (1978). Soil mapping at 1:1,000 scale revealed characteristic soil-regolith-slope relationships within first order drainage basins. The occurrence of the major soil groups (yellow-brown earths, podzols, gley podzols and gleys) is primarily controlled by slope position (nose, sideslope or hollow) and hillslope hydrology (catenary relationships).

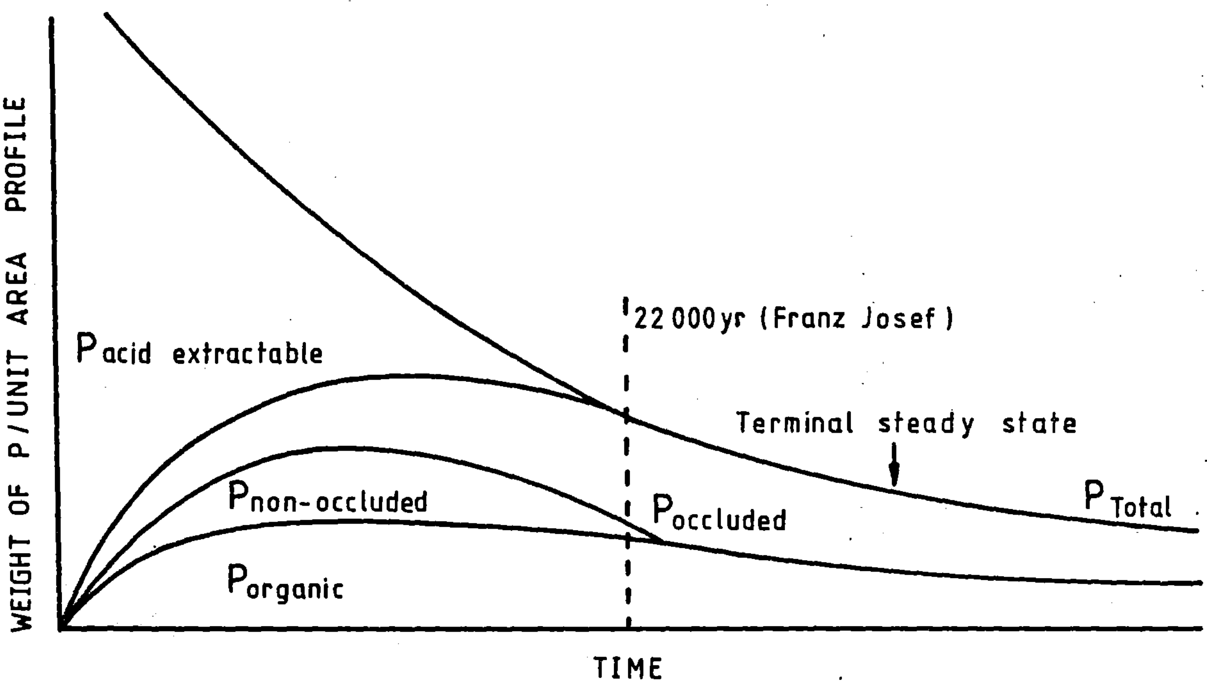


Fig. 6: Changes in forms and amounts of soil phosphorus with time (from Walker and Syers, 1976).

Soil chemical and mineralogical characteristics also varied with topography. Simmons (1982) describes a catena developed on dissected high terraces near Hari Hari. Variation in soils (from gley podzols to yellow-brown earths, organic soils and peaty gley podzols) was observed with minor changes in topography.

2.5 EROSION - ASSESSMENT TECHNIQUES AND TERMINOLOGY

In this section methods for assessing the type and activity of erosion processes, over time scales ranging from 10^0 to 10^5 years, are briefly reviewed and terminology for the description of present day forms and processes of erosion is introduced.

2.5.1 Erosion assessment techniques

Assessment of soil erosion involves investigation of:

(a) the types of erosion processes operating or a surrogate measure, the forms of erosion occurring;

(b) the amounts of material involved in the various processes;

(c) the rates of operation of the various processes.

Methods for assessing the type and activity of erosion processes include direct (inductive) and indirect (deductive) techniques. In the inductive approach general principles are inferred from detailed studies of specific instances while the deductive approach infers from the general case to the specific. In any assessment programme it is desirable to use a combination of techniques.

Direct measurements of specific processes have been undertaken in both the field and laboratory. Field examples include measurement of soil creep (Owens, 1969; Young, 1978) surficial rock debris movement on talus slopes (Gardner, 1969, 1979; Luckman, 1978), solifluction (Washburn, 1973), and soil loss (Hayward, 1968, 1969; De Ploey, 1977; Bovis and Thorn, 1981). There have been attempts to compare and evaluate the array of erosion processes operating in mountainous environments (Rapp, 1960; Iveranova, 1969). Numerous erosion processes have been investigated by laboratory experiments including freeze-thaw action (Corte, 1962; van Steijn, 1977), soil creep (Kirkby, 1967), and rainsplash/sheetwash erosion (Mucher and De Ploey, 1974; Moeyersons and De Ploey, 1976). Also included in this inductive approach are empirical methods such as the Universal Soil Loss Equation (Wischmeier and Smith, 1965, 1978), used to predict soil loss from sheet and rill erosion caused by Hortonian overland flow.

While inductive techniques are appropriate for some purposes they have inherent limitations. Temporal and spatial variability of erosion processes, and variations between processes, create major sampling problems. The slowness of operation of most processes means that long-term observation programmes are necessary. Instrumentation problems, including interference with the process being measured are a practical difficulty. In estimating the importance of various processes, information is needed on the magnitude of the process, its frequency of operation and its effectiveness in shaping the landscape (Wolman and Miller, 1960; Wolman and Gerson, 1978). This involves consideration of the importance of rare, high magnitude events as opposed to frequent, low magnitude events and the effect of threshholds.

Because of these limitations deductive techniques have commonly been used to assess erosion. The sediment load in streams has been widely used to calculate rates of erosion (e.g. Langbein and Schumm, 1958; Fournier, 1960; Judson and Ritter, 1964; Adams, 1978, 1980a; Griffiths, 1979b, 1981; Griffiths and McSaveney, 1983a) as has the amount of sediment accumulated in resevoirs over a known time period (e.g. Langbein and Schumm, 1958; Eardley, 1966; Thompson, 1976; Bishop *et al*, 1984). These measurements integrate the products of many erosion processes and only estimate the amount of debris leaving a drainage basin. They do not take account of the changes in sediment storage within a basin. There are also a number of potential sources of error that may be important, including methods of data collection, human interference in recent times and its impact on sediment yields, assessment of bedload and dissolved load components of total yield (Meade, 1969; Walling, 1978). However sediment yield has been widely used to provide estimates of regional rates of erosion.

On a more detailed scale direct observations and measurements of form have been used to infer the nature and rate of operation of processes, particularly mass movement forms of erosion. This approach is useful because many processes of mass movement are rarely observed due to their episodic nature. Examples include studies of denudation resulting from heavy rainfalls in the Scandanavian mountains (Rapp, 1963; Rapp and Stromquist, 1976; Larsson, 1982) and Himalayas (Starkel, 1972) and from an earthquake in Papua New Guinea (Pain and Bowler, 1973); studies of soil avalanching in Hawaii (Scott and Street, 1976) and Japan (Tanaka, 1976); and studies of the relationship between landslip morphometry and process (Crozier, 1973; Blong, 1973). Difficulties with this approach include the need to assume unique relationships between form and process, the multiplicity of processes that occur at one time to produce particular forms, and difficulties in extrapolating short-term, often catastrophic, rates over longer time periods.

In order to circumvent the slowness and episodic operation of erosion processes, and the long time periods over which they operate, evolutionary methods (or space-time substitution) have been used in suitable areas. Where a land surface, or succession of land surfaces can be dated accurately the modification of that land surface can be evaluated quantitatively. Savigear (1952) examines change in a sequence of slope profiles along the Welsh coastline that have been beyond the influence of coastal erosion for varying periods of time. Erosion rates, based on the modification of lava sheets of known age, are calculated for areas in Papua New Guinea (Ruxton and McDougall, 1967) and California (Marchand, 1971). Measurements of the amount of material removed from around tree roots, combined with estimates of tree age are used to estimate rates of soil loss (Eardley and Viavant, 1967; La Marche, 1968; Dunne *et al*, 1978). Studies of this type provide a medium- to long-term perspective on erosion rates and are particularly valuable when combined with estimates of current rates of erosion from process studies (Imeson *et al*, 1980).

Also included in the evolutionary approach is the interpretation of the stratigraphic record, and in particular soil stratigraphy (see section 2.3). Geomorphic interpretations of the stratigraphic record document process variability over long time periods (10^2-10^5 years), and responses of geomorphic systems to events of varying cause and magnitude. Such interpretations are best made where the record is most complete, direct numerical dating is available, ties between geomorphic systems and depositional environments can be made, and independent evidence on extrinsic controlling factors is available (Johnson, 1982). Thorn (1982) suggests that soil geomorphology is a potential key in linking process studies and stratigraphy because it integrates the effects of currently operative processes and landscape history.

2.5.2 Terminology

Classifications of erosion consider both form (morphology at a given point in time) and process (the agents active in causing form to change). Inferences about process have commonly been made from observations of form, although it is acknowledged that recognising the process from the debris it leaves behind is not always easy (Selby, 1982). Further difficulty arises from the need to assume relationships between form and process when it is recognised that different processes may give rise to similar forms, and the same process, through its variability in time and space, may result in different forms (Embleton and Thornes, 1979). However, in view of the difficulties in directly observing process (2.5.1) the deductive approach is most commonly used.

Erosion is defined here as the various ways in which rock and soil debris is detached and removed, and is distinguished from weathering which is confined to the physical disintegration and chemical decomposition of rock in place. Weathering, of which soil development is a part, is a relatively static process that commonly is the forerunner to failure and movement. This distinction between erosion and weathering is an artificial division of a continuum of processes involved in landform modification.

Several classifications of the types of erosion that affect the earth's surface have been proposed which consider both form and process (the same terms are commonly used for both). No comprehensive system is as yet universally acceptable, since any classification imposes arbitrary boundaries useful for subdividing a continuum of forms.

Forms of erosion can be distinguished initially on the basis of the dominant transporting agency. The main agents of erosion on slopes are water (fluvial erosion), gravity (mass movement), wind (aeolian erosion), and ice (glacial erosion). This discussion is limited to fluvial, mass movement and aeolian erosion since they are the main forms of contemporary slope erosion in formerly glaciated areas of the Southern Alps. Fluvial and aeolian erosion are characterised by movement of particles as individuals, while in mass movements neighbouring particles move as a coherent mass. Many erosion forms result from combinations of the above agencies.

Little difficulty is experienced in applying terminology to pure fluvial and aeolian erosion forms. Fluvial erosion on hillslopes is recognised from the occurrence of sheet wash, rills, gullies, subsurface pipes and the development of high drainage density. Sheet wash is the only type that may be difficult to identify, since bare areas characteristic of this form of erosion are also formed by aeolian erosion. Rainsplash erosion, a gravity-induced fluvial form of erosion, results from the impact of raindrops on the ground surface.

It is recognised by the development of features such as stone capped pinnacles. Aeolian erosion is commonly recognised from the characteristic depositional features it produces (e.g. dunes). It is most important in dry or seasonally dry environments.

Máss movement comprises all gravity-induced movements, without the direct aid of other agencies (Hutchinson, 1968). Water and ice, however, are frequently involved as contributing factors to mass movement. A range of materials and processes are involved in mass movement, with a consequent variety of resultant forms.

Classifications of mass movements are contained in Sharpe (1938), Campbell (1951), Varnes (1958, 1978), Hutchinson (1968), Carson and Kirkby (1972), Northey *et al* (1974), and Selby (1982). Distinguishing between different forms of mass movement requires consideration of the material involved (rock, regolith, soil), the mode of deformation, the velocity and mechanism of movement, geometry and water content of the moving mass. Carson and Kirkby (1972) use a triangular diagram to show the three main types of movement, and to indicate the rate of movement and water content (Fig. 7). They distinguish the following types of movement:

(a) Slide - with a sharply defined failure surface. Can be further subdivided according to whether the failure surface is planar (translational slide) or curved (rotational slide). Other authors have used a fourth type of movement, falls, which are a special case of slides where the failure plane is near vertical;

(b) Flow - shear is distributed throughout the moving mass with no sharply defined failure surface;

(c) Heave - soil expands perpendicular to the surface and subsequently contracts leading to slight downslope movement.

Actual mass movement processes are rarely, if ever, pure slide, flow or heave, but are a combination of all three and

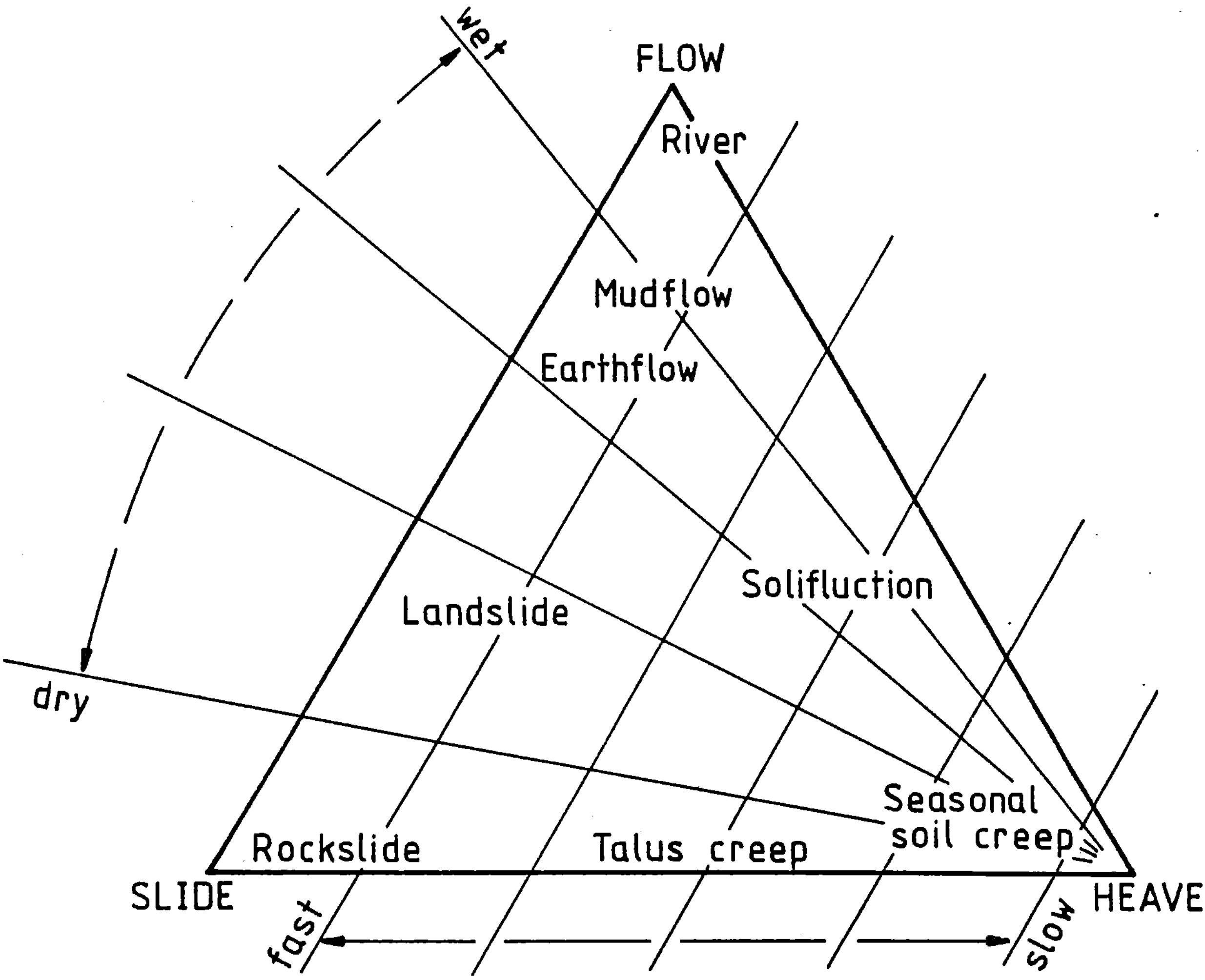


FIG. 7: Classification of mass movement processes (from Carson and Kirkby, 1972).

therefore a triangular diagram provides a convenient basis for conceptually classifying the types of movement involved in different processes, or in producing different forms. Further subdivision of mass movements is made on the basis of the material involved in failure, whether it be bedrock, unconsolidated sediments, or regolith/soil. Many failures however include a range of materials and therefore terminology is not easily applied. It is not proposed to present any comprehensive classification of mass movement erosion, since that is beyond the scope of this thesis. Instead specific terms are used to describe the forms of erosion observed in Cropp basin. Terminology follows Varnes' (1958, 1978) description of mass movement (Fig. 8), and Davies (1972) description of frozen ground and nivation phenomena.

The following terms are used to refer to the material involved in movement. Bedrock or rock designates relatively consolidated and unweathered material that was in stratigraphic and lithologic sequence before the initiation of movement. Regolith is defined as including the soil and all unconsolidated weathered debris of whatever geological origin, and weathered bedrock (Tonkin, 1984). Regolith may be further subdivided into debris mantle regolith (upper part of the regolith comprising the soil and all unconsolidated, oxidised or reduced surface sediments of residual or transported origin) and the bedrock regolith (lower part of regolith that has been altered by subaerial weathering processes but that maintains a structure and stratigraphy comparable to the unweathered bedrock). The distinction between debris mantle and bedrock regoliths is useful in areas of long term stability and obvious deep weathering. In young landscapes underlain by highly fissile, and often fractured, rocks it is impossible to meaningfully distinguish between bedrock and bedrock regolith, and therefore the latter term is not used. The term debris is used for regolith that contains a significant proportion of coarse material (20 to 80% greater than 2 mm according to Varnes, 1978) and is distinguished from earth (with 80% of fragments less than 2 mm; Varnes, 1978).

<table>
<tr><td>TYPE OF</td><td colspan="4">TYPE OF MATERIAL</td></tr>
<tr><td>MOVEMENT</td><td colspan="2">Bedrock</td><td colspan="2">Engineering soils</td></tr>
<tr><td>Falls</td><td colspan="2">ROCKFALL</td><td colspan="2">DEBRIS/EARTH FALL</td></tr>
<tr><td>Topples</td><td colspan="2">ROCK TOPPLE</td><td colspan="2">DEBRIS/EARTH TOPPLE</td></tr>
<tr><td rowspan="2">Few units
Slides
Many units</td><td>planar
ROCK BLOCK SLIDE</td><td rowspan="2">rotational
ROCK SLUMP</td><td>planar
DEBRIS/EARTH BLOCK SLIDE</td><td>rotational
DEBRIS/EARTH SLUMP</td></tr>
<tr><td>ROCK SLIDE</td><td>DEBRIS/EARTH SLIDE</td><td>FAILURE BY LATERAL SPREADING</td></tr>
<tr><td rowspan="3">Flows
Increasing Water Content ↓</td><td colspan="4">decreasing particle size ⟶</td></tr>
<tr><td colspan="2" rowspan="2">ROCK AVALANCHE</td><td colspan="2">SOIL CREEP</td></tr>
<tr><td>DEBRIS AVALANCHE

DEBRIS FLOW</td><td>EARTHFLOW

MUDFLOW</td></tr>
<tr><td>Complex</td><td colspan="4">Combination of two or more principle types of movement</td></tr>
</table>

FIG. 8: Classification of mass movements (after Varnes, 1958, 1978).

Debris and earth together constitute soil, in an engineering sense, which is equivalent to debris mantle regolith as previously defined. Colluvium is used for debris mantle regolith formed on hillslopes.

The processes operating to modify hillslopes, and their resultant erosion forms are important, but in considering landscape development a more important distinction is based on the immediate source of moving debris (Carson and Kirkby, 1972). If transport processes are more rapid than weathering, only a thin regolith is able to develop and movement is said to be 'weathering limited'. If weathering rates are more rapid than transport processes, a thick regolith develops and movement is said to be 'transport limited'. These concepts express the relationships between surficial weathering processes, soil development and slope development. There is a general association between weathering limited slopes, relatively thin regolith and soils, the existence of important threshold slope angles and parallel retreat at these angles; and between transport limited slopes, thick regolith and soils, and convex-concave slopes which become progressively less steep without significant threshold slope angles (Carson and Kirkby, 1972).

CHAPTER 3 - INTRODUCTION TO THE STUDY

3.1 SCOPE OF THE STUDY

Studies of geomorphology, contemporary forms of erosion, and the development and distribution of soils, are used to infer the history of soil erosion, and to assess its effect on the land surface in Cropp basin. The relative development of soils is used to indicate land surface age. The spatial arrangement and chronological array of soils provides a record of the magnitude and frequency of erosional episodes in Cropp basin. No direct observations of erosion processes were made during this study. Observations of contemporary erosion forms, and their distribution, combined with studies of past events and soil-landscape relationships are used to deduce a history of landscape evolution, assuming that the present is the key to the past (Principle of Uniformitarianism; Thornbury, 1969, pp 16-17).

The interpretation of relative soil development and the soil pattern requires a knowledge of the sequence of changes of soil physical and chemical properties with time, together with the lateral changes within the soil continuum exhibited at any given development stage. This has been provided by the study of soil chronosequences (dated independently by stratigraphy, dendrochronology and radiocarbon dating) and limited study of catenary relationships. Selected soil chemical analyses were undertaken to confirm the interpretation of these sequences, and to document soil properties characteristic of this environment. Study of the plant communities associated with the soil sequences provides an understanding of plant succession following disturbance of vegetation by erosion.

Firstly, the environmental characteristics of the study area that are important in understanding erosion processes and soil development are outlined.

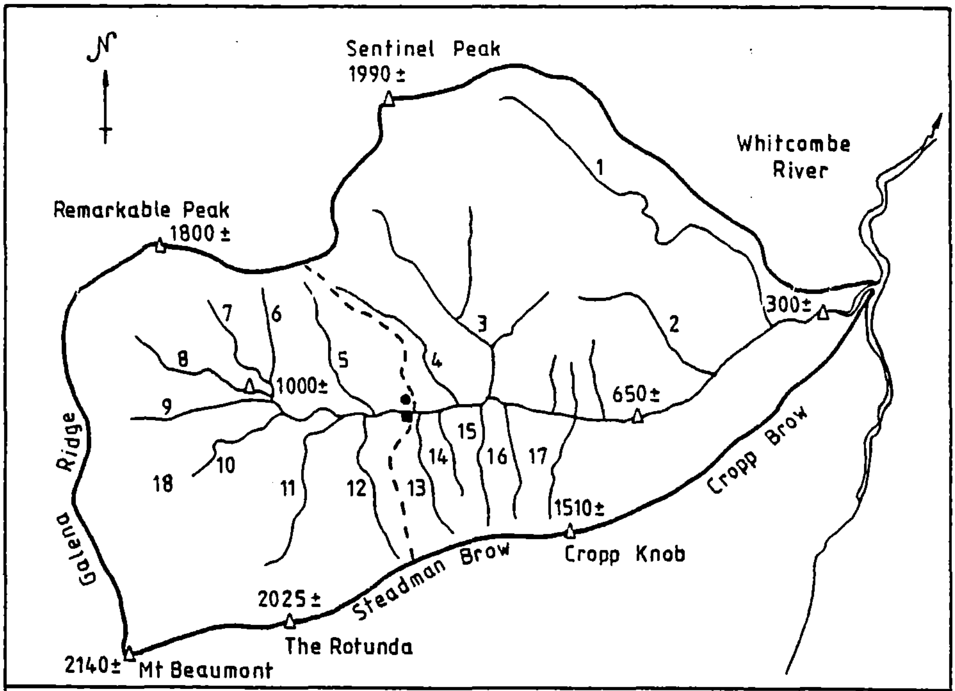
3.2 LOCATION AND PHYSIOGRAPHY

Cropp River drains a 28.5 km^2 basin into Whitcombe River, a major tributary of the Hokitika River (Fig. 2). Situated in the western front ranges of the Southern Alps (43°05'S, 171°00'E), the basin has a predominant east-west orientation and is nearly rectangular (7.5 km long by 3.5 km wide) in plan (Fig. 9). Elevations range from 2140 m at Mt Beaumont to 240 m at the confluence of the Cropp and Whitcombe Rivers.

Cropp River is fed by a series of steep headwater streams (Fig. 9). The source is regarded as the stream draining from between Remarkable Peak and Galena Ridge, which is progresively joined by streams draining Remarkable Peak, Galena Ridge, Beaumont Basin* and Mt Beaumont. The river then flows through a flatter valley, puncutated by gorges, and is joined by major tributary streams from the north and smaller streams from the south. Between Snowy Stream and Danger Gully is a prominent bedrock knob (Tarkus Knob*) and the most extensive area of gently sloping topography in the basin (Hut Flat*). Below Reckless Torrent and Cream Creek* Cropp River enters a narrow gorge that falls steeply to the Whitcombe River. Noisy Creek, a hanging valley formed behind resistant bedrock, joins the Cropp near the Whitcombe end of this lower gorge.

Physiographically the basin is characterised by high drainage density and extremely steep slopes (40-60°+ commonly) that are generally rectilinear in profile. Stream grades are steep and uneven. Weakly modified U-shaped cirques are present at the heads of the larger tributaries but most of the glacial landscape has been modified by post-glacial fluvial and mass movement erosion. Valley profiles are generally V-shaped (see Chapter 4 for discussion of geomorphology). The landscape is typical of the western geomorphological region of Whitehouse (1985a).

*Informal names (see Fig. 9)



KEY

-----	Boundary to upper Cropp basin
■	Cropp Hut
(line)	Major stream
Δ	Topographic elevation in metres
1	Noisy Creek
2	Pounamu Creek *
3	Reckless Torrent
4	Snowy Stream
5	Danger Gully
6	Warning Creek
7	Remarkable Creek *
8	Cropp River
9	Steep Creek *
10	Waterfall Creek *
11	Rotunda Creek *
12	Whinging Creek *
13	Hut Creek *
14	Fan Creek *
15	Steadman Creek *
16	Cream Creek *
17	Knob Creek *
18	Beaumont Basin *
●	Tarkus Knob *

SCALE

0 1 2 3 4km

* Informal names

FIG. 9: Map of Cropp basin, drawn from NZMS 1 sheets S64 and S65 (boundaries of Cropp basin are enclosed within the following grid coordinates: East - 485 to 575; North - 090 to 155).

3.3 CLIMATE

Cropp River lies within a narrow zone along the western front ranges of the Southern Alps characterised by frequent heavy rainfalls (> 100 mm/day) and annual precipitation in excess of 10,000 mm. Characteristics of precipitation, temperature and wind, recorded mainly at Cropp Hut (Fig. 9), are briefly described.

3.3.1 Precipitation (summarised from Griffiths and McSaveney, 1983a).

Air masses, moving west off the Tasman Sea, ascend the Southern Alps releasing moisture as a result of adiabatic cooling. This results in a narrow zone of extremely high annual precipitation (10,000 to 12,000 mm) in the western front ranges of the Southern Alps (Fig. 10a). Rain is heaviest when derived from frontal disturbances within intense extra-tropical cyclones advecting very moist subtropical air from the northern Tasman Sea. A typical weather pattern causing high rainfall in Cropp basin is illustrated in Fig. 10b.

Basin mean annual rainfall is 10,800 mm a^{-1} with a recorded range from 8,800 mm a^{-1} at Cropp-Whitcombe confluence (274 m above sea level) to 11,800 mm a^{-1} near the head of the basin (1524 m a.s.l.). An isohyetal map of the basin showing the location of raingauges is given in Fig. 11. Statistics derived from measurements made by an automatic pluviometer at Cropp Hut are given in Table 1a. There are more than 200 rain days per annum and substantial rain (> 25 mm) falls on more than half the days in the year. A daily total greater than 100 mm can be expected at least once each month, with a maximum recorded 24 hr rainfall of 555 mm. Monthly totals are quite variable because of the influence of large storms, but usually exceed 500 mm, and commonly exceed 1000 mm (maximum 1819 mm). Precipitation is well distributed throughout the year although totals in June and July are commonly lower (less than 500 mm). Rainfall intensities in excess of 25 mm

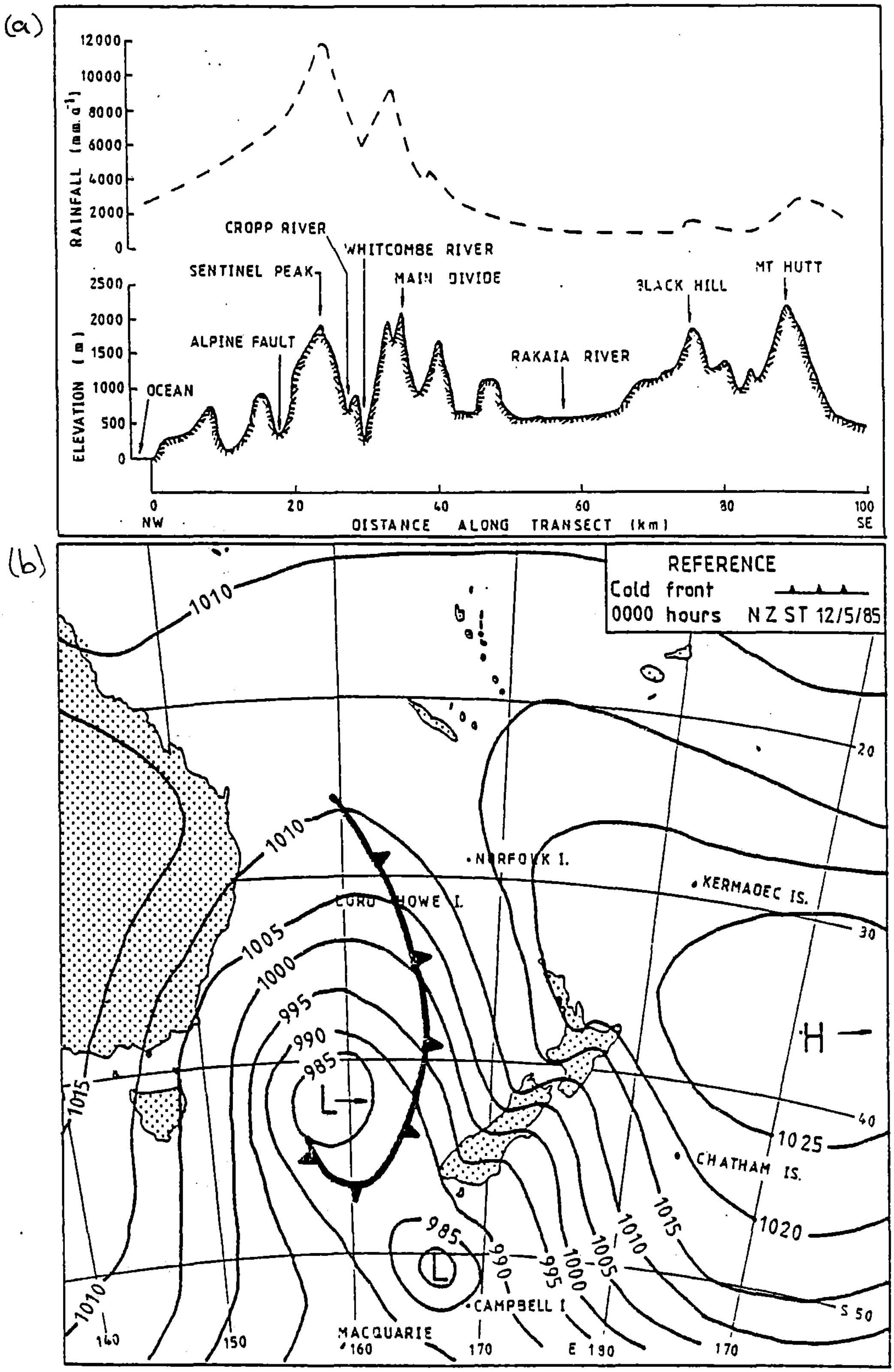


FIG. 10: a) Distribution of rainfall in a transect across Southern Alps including Cropp River (from Griffiths and McSaveney, 1983b).

b) Typical weather pattern causing high rainfall in Cropp basin.

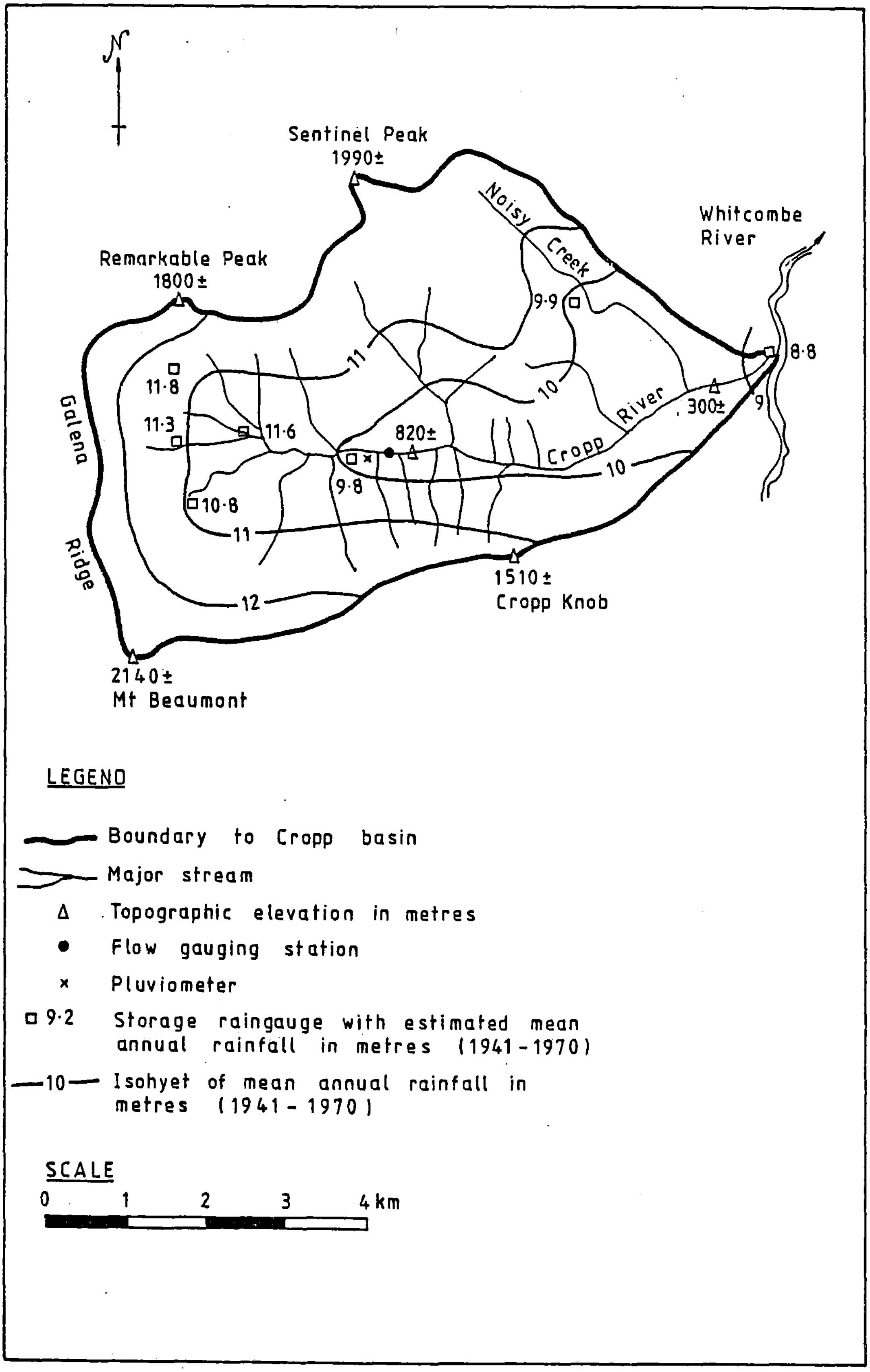


FIG. 11: Isohyetal map of Cropp basin (from Griffiths and McSaveney, 1983a).

a) Rainfall totals

Year	Mean daily (mm)	Max daily (mm)	Monthly		Annual total (mm)	No. of rain days	Max. period		No. of days greater than		
			Minimum (mm)	Maximum (mm)			wet (days)	dry (days)	100 mm	200 mm	300 mm
1980	25.9	248	291	1670	9440	212	25	19	35	5	-
1981	29.2	428	107	1620	10700	219	17	12	36	9	2
1982	28.8	490	250	1523	10500	205	12	17	35	8	3
1983	34.4	399	183	1569	12550	234	19	10	43	14	4
1984	32.7	338	502	1819	11950	212	20	10	41	15	2

b) Rainfall-duration-return period relations (Whitehouse, 1985b).

Return period (yr)	Rainfall depths (mm) for stated durations			
	10 min	1 hr	6 hr	24 hr
1.1	13.0 $\pm$ 1.1	53 $\pm$ 4	198 $\pm$ 16	444 $\pm$ 34
2.33	14.6 $\pm$ 1.4	61 $\pm$ 5	226 $\pm$ 20	509 $\pm$ 43
5	16.2 $\pm$ 1.7	68 $\pm$ 7	252 $\pm$ 25	572 $\pm$ 52

Recorded maximum rainfalls (mm)			
6 min	1 hr	6 hr	24 hr
18.5	67.5	287.5	555

TABLE 1: Statistics of rainfalls measured at Cropp Hutt (elevation = 865 m).

hr^{-1} occur frequently, with a maximum recorded 6 minute intensity of 18.5 mm. Rainfall-duration-return period relations are presented in Table 1b.

The contribution of snow to annual precipitation has not been assessed, but is probably 10-20% of basin annual precipitation (M.J. McSaveney, pers. comm. 1984), since at nearby Ivory Glacier, at 1400-2080 m altitude, snow contributes 25% of annual precipitation (Anderton and Chinn, 1978). Duration of snow cover increases with altitude and is about 3 months at timberline. Small areas of permanent snow exist at the highest elevations. Snow commonly falls in the forest and scrub covered areas, but generally does not persist for more than a few days.

Frequent intense rainfalls and a high annual total are the dominant influence on erosion processes, geomorphology and soil development and distribution in Cropp basin. Snow cover is an important influence on vegetation patterns.

3.3.2 Temperature

Air temperature at 1.3 m and soil temperature at 0.5 m depth were measured at Cropp Hut (865 m a.s.l.) with temperature probes attached to automatic Leopold and Stevens digital recorders. Hourly values of soil temperature and half hourly values of air temperature were recorded.

Mean annual air temperature is estimated to be 5.5 ± 0.5°C. Regular seasonal and diurnal fluctuations are evident which are modified by storms. January and February tend to be the warmest months with June and July the coldest. Mean daily air temperature ranges from 15° to -5.5°C. The recorded temperature range is from 23.4°C, to -10.1°C. Frosts may occur at any time of the year, but are most common in June and July, occurring on up to 26 days in a month. Temperature statistics for those months with a complete record are presented in Fig. 12.

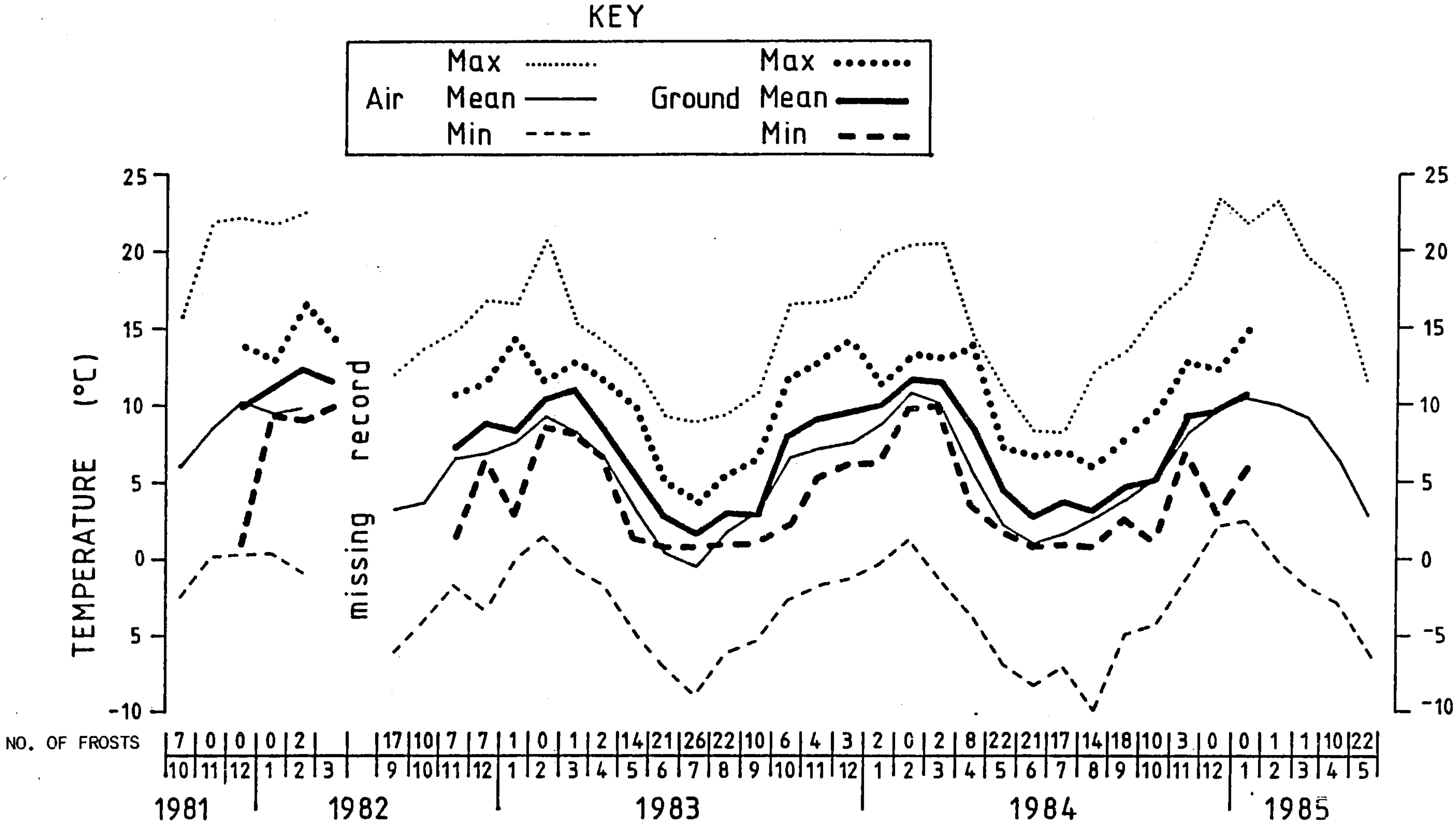


FIG. 12: Comparison plot of monthly mean, maximum and minimum ground (0.5 m depth) and air (1.3 m) temperatures.

At 0.5 m soil depth there is a restricted range, and less fluctuation compared to air temperatures. Seasonal patterns are similar for both air and soil temperature. Mean annual soil temperature is estimated to be 6.0 $\pm$ 0.5°C. Daily means range from 14.7°C to 1.8°C. Maximum recorded soil temperature is 16.4°C and minimum is 0.6°C. At 0.5 m depth the soil was never frozen during the period of record, however freezing and thawing are relatively common nearer the surface. Temperature statistics for those months with a complete record are presented in Fig. 12.

3.3.3 Wind

Windrun was recorded by a monthly Lambrecht anemograph at a height of 3 m above the ground. For the period March 1980 to May 1984 a 60% record of windrun was obtained with gaps in the record well distributed seasonally. The record is therefore considered representative of the windrun characteristics of the Cropp Hut site.

For the period of record, mean daily windrun was 119 km with a maximum of 503 km. Highest frequency of daily windrun is in the range 50-200 km. Windrun tends to be at a minimum in mid-winter and at a maximum in late spring. Maximum gusts have not been recorded, but at nearby Ivory Glacier gusts in excess of 200 km hr^{-1} have been recorded (Chinn, 1980). Similar gusts are likely in exposed areas at high elevations and probably also at Cropp Hut. Cropp valley is calmer than most valleys east of the Main Divide (Chinn, 1980), but very strong wind gusts are a notable feature of the climate.

Wind direction is dominantly west to north-west. Gale force winds are also likely from the south-east. Wind direction is modified locally however, and at Cropp Hut is generally up or down the valley, with frequent reversals.

3.4 HYDROLOGY AND SEDIMENT YIELD

Griffiths and McSaveney (1983a) describe the hydrological characteristics of upper Cropp basin (Fig. 9) for the

1980-1982 period. Cropp River is a mountain torrent with a bed grade of 0.1 or more in the headwaters, to 0.044 at the outlet to upper Cropp Basin. The channel features both flat-and-rapid and pool-and-riffle channel geometries, defined by channel boundaries of bedrock or boulders and cobbles (Wentworth scale). Drainage density, D (Scheidegger, 1970), is very high. If only major channels are considered D = 5 $km.km^{-2}$, but there also exists a myriad of micro-channels, with perennial and ephemeral flow, which indicate that D is at least an order of magnitude greater. During storms the channel network expands enormously.

Flooding in upper Cropp basin is characterised by a high frequency of events, rapid runoff response, and a high ratio of runoff to rainfall leading to rapidly peaking floods. Statistics for flows and runoffs are presented in Table 2. Mean annual flood is estimated at 370 $\pm$ 30 $m^3\ s^{-1}$. The maximum instantaneous discharge recorded is estimated at 524 $m^3\ s^{-1}$. Simple and Hortonian overland flow, and subsurface flow (Dunne, 1978) are thought to dominate streamflow generation. Prompt supply of streamflow to a very dense network of surface channels, and the presence of subsurface pipes (with dimensions up to 30 x 20 cm), yields rapid storm runoff under short-duration high-intensity rainfall. There is a nearly constant minimum time (24 $\pm$ 2 minutes) for rise of a flood resulting from short-duration rainfall excess. More than half the runoff from storm rainfalls leaves the basin within an hour of rainfall ceasing, and half the remaining storm runoff depletes within about 2 days.

Only the suspended component of river transported load has been measured due to practical limitations. Measured specific annual suspended sediment yield is 29,600 $\pm$ 2,500 t $km^{-2}\ a^{-1}$ representing a basin denudation rate of 11 $\pm$ 1 mm a^{-1}, assuming a bedrock density of 2.65 t m^{-3}. Bedload yield is surmised to contribute less than 10% of total load. Sediment is supplied at numerous locations along the extensive channel network. Fluvial and mass movement erosion forms are the main sediment supplying mechanisms.

		Year 1980	Year 1981
FLOWS (m^3s^{-1})	Mean	4.38	4.57
	Standard deviation	$\pm$ 6.44	$\pm$ 7.04
	Median	2.60	2.63
	Instantaneous - Min	0.677	0.565
	- Max	524$\pm$	132
RUNOFFS	Annual runoff (mm)	11,300	11,800
	Max instantaneous specific yield ($m^3s^{-1}km^{-2}$)	43.0	10.8
	Peak runoff (mm h^{-1})	155$\pm$	39

TABLE 2: Statistics of flows and runoffs measured at Cropp at Gorge flow gauging station (from Griffiths and McSaveney, 1983a).

3.5 GEOLOGY

A relatively detailed account of the geology of Cropp River is given (summarised from Hawkes, 1981) since geologic factors have a marked influence on geomorphology and erosion.

3.5.1 Bedrock lithologies

Cropp River lies within the Alpine schist belt of the axial ranges of the Southern Alps and is underlain by high grade metamorphic rocks of greenschist to amphibolite facies. Most of the schists are quartzofeldspathic and compositionally similar to nearby, less metamorphosed Torlesse Group rocks. In addition there are metavolcanics, metacherts and marble in an apparently simple stratigraphic succession.

Six distinctive lithologic groups occur:

(a) Metapsammite

Coarser grained (0.1 - 1 mm) schistose, quartzofeldspathic rocks that form the major bedrock component. Major constituents are quartz and feldspar with chlorite, biotite, epidote and garnet developed in varying proportions depending on metamorphic grade. Mean grain size is silt or fine sand.

(b) Metapelites

These are the fine grained equivalents of metapsammites. Grain sizes are very fine (.05 mm $\pm$) and colours are darker than for psammites. Compositionally they are similar to psammites. There is a complete gradation between these two rock types.

(c) Metacherts

Fine textured milky white rocks that often occur in discrete beds interbedded with metapelites. They are dominantly quartz with 10 to 30% albite and minor micas and garnets.

(d) Metavolcanics

Two classes are represented:

(i) Greenschists - occur concordant with other lithologies in quartzofeldspathic schists. They are texturally similar to metapelites but have a greyish green colouration due to the abundance of chlorite. Quartz, feldspar and calcite are the other major constituents. They are volumetrically important in the upper biotite zone comprising 20-40% of rocks exposed.

(ii) Ultramafics - two well exposed, elongate pods of serpentine-dunite and serpentinite surrounded by envelopes of talc-magnesite and greenschist are present in the lower Cropp-Noisy Creek area. These rocks belong to the Pounamu formation (Morgan, 1908; (Koons, 1978; Cooper and Reay, 1983).

(e) Marble

Occurs in minor proportions associated with metapelites and metavolcanics.

Numerous gradations between these general groups are present. The rocks represent a metamorphosed stack of Torlesse-like sandstones, mudstones and cherts interspersed with basic volcanics and/or volcaniclastics. Distribution of lithologies is shown in Fig. 13.

3.5.2 Metamorphic zonation

Metamorphism involved 3 temperature maxima which varied in extent and area, thus complicating textural zonation of the rocks. In terms of metamorphic facies (defined by characteristic mineral assemblages) the rocks belong to the greenschist and amphibolite facies. Chlorite, biotite and garnet zone rocks of the greenschist facies are present and oligoclase zone rocks of the amphibolite facies are present (see Fig. 13).

Textural zonation (defined in terms of rock fabric) is complicated by isolated narrow zones of high textural grade superimposed on an overall pattern of increasing grade towards the Alpine Fault. Thus there is a complex relationship between textural and mineralogical zones. The overall pattern involves three distinctive belts: an eastern area with mineralogically high grade but texturally poorly developed

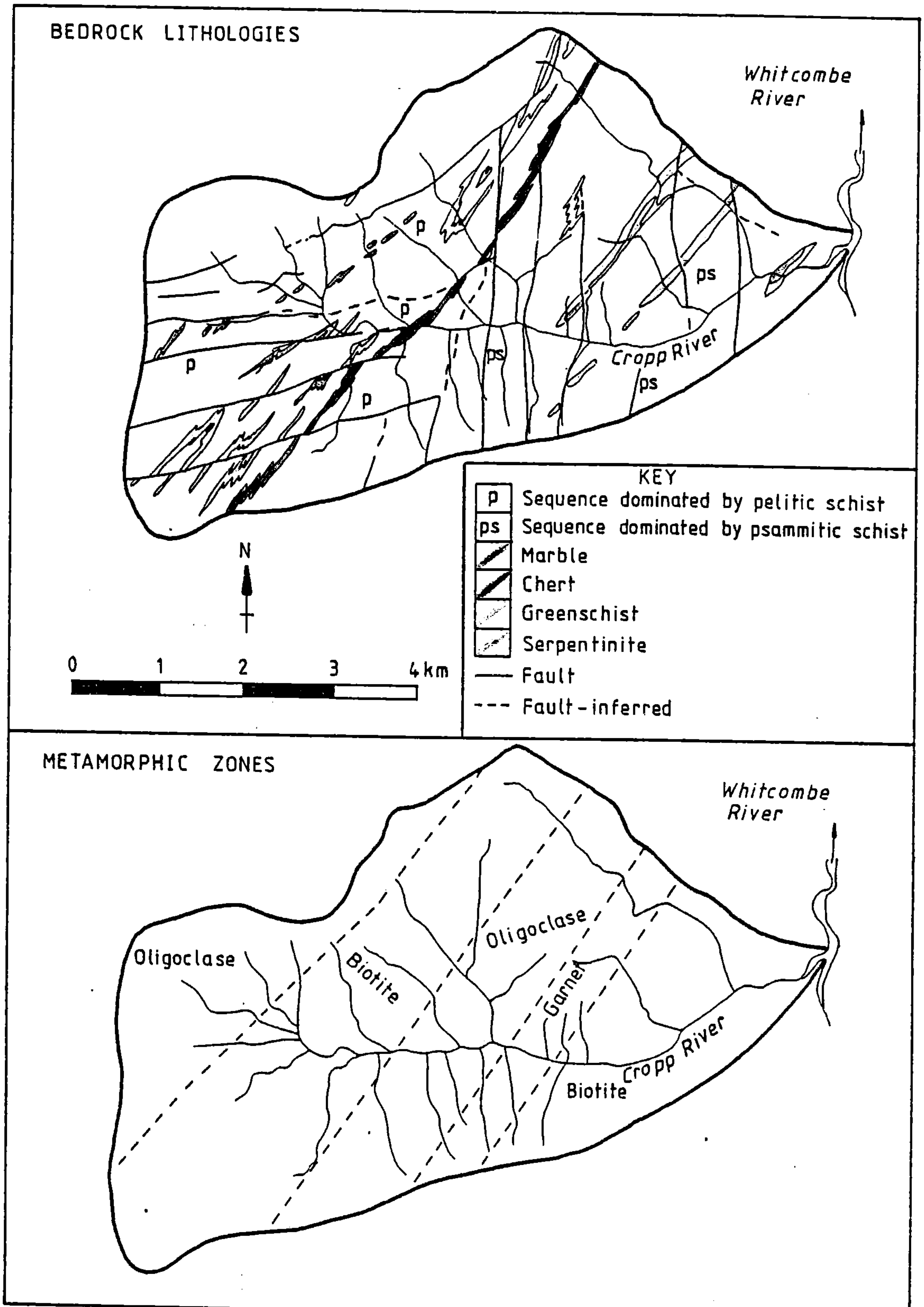


FIG. 13: Simplified geologic map of Cropp basin (from Hawkes, 1981).

rocks; a western zone of mineralogically and texturally high grade schists; and an area of mineralogically and texturally low grade rocks dividing these zones.

3.5.3 Structural patterns

Four deformation episodes are recognised, all involving folding. The first three episodes were associated with metamorphic events and development of metamorphic fabric. The last, involving minor folding, is associated with a conspicuous ENE-trending, steeply dipping, dextral fault set and consists of small scale, moderately plunging kink folds that are post-metamorphic. Probably of the same age is a set of sinistral, north trending, steeply dipping faults that forms a conjugate fault system with the ENE-trending faults (Fig. 13). A late stage southwest dipping joint set is prominent on the north side of Cropp valley. These joints relate to gravity collapse, and are a major control on geomorphology (Fig. 14). In general terms the rocks are highly fissile and commonly intensely fractured.

3.5.4 Uplift rates and seismicity

Uplift of the Southern Alps is the result of crustal shortening due to obliquely convergent motion between the Indian and Pacific plates (Walcott, 1984). The shortening has resulted in mountain building where continental rocks on each plate abut. Horizontal and vertical movement is distributed across the Southern Alps (Adams 1978, 1980a; Wellman, 1979). Maximum crustal shortening, and hence uplift, occurs southeast of the Alpine Fault (Wellman, 1979). Whilst there is a long history of activity along the plate margin, uplift which formed the present Southern Alps, began in the late Miocene (10 million years ago) and still continues (Walcott, 1978; Adams, 1979).

Hawkes (1981) infers an uplift rate for Cropp River from K/Ar dating of rock samples, estimates of geothermal gradient and geomorphic considerations. He suggests that uplift is occurring at present localised on the Alpine Fault in the

FIG. 14: Photo of upper Cropp basin. South facing slopes (right of photo) are underlain by pelitic schist and are highly dissected as a result of gravity collapse on a southwest dipping joint set. Linear channels of streams draining Galena Ridge, at head of valley, are controlled by ENE-trending faults.

west, and on northeast-trending faults in the east. Cropp River may be in the zone of maximum uplift with an estimated rate of 12 $\pm$ 2 $mm.a^{-1}$. This estimate is consistent with the estimates of Wellman (1979) (Fig. 15).

Cropp River lies within the central seismic region of Eiby and Reilly (1976). This region has little current shallow seismic activity, requiring either the expectation of a large event or that deformation is largely aseismic (Reyners, 1984). The former is considered more likely with recurrence intervals of about 500 years inferred for major fault displacements on the Alpine Fault (Adams, 1980b; Berryman, 1984). A rock avalanche in the lower Cropp (Fig. 16) may have been earthquake triggered since Wardle (1980) notes similar features associated with the Alpine Fault.

3.6 VEGETATION

Descriptions of the plant communities in the Hokitika and Whitcombe drainage basins are contained in Wardle (1960), Wraight (1960), Holloway (1966), and James *et al* (1973). Also of relevance are descriptions of plant communities in Westland National Park given by Wardle (1977, 1980), although there are some significant differences between the two areas. The following is a general account of the major vegetation associations. A detailed description of vegetation patterns in relation to observed soil patterns is given in Chapter 7.

At Cropp River vegetation can be divided into associations based on altitudinal zonation: low altitude and montane forest, subalpine forest and scrub, alpine grasslands and alpine barrens. These broad zonations apply to large areas. In detail, however, species composition can be extremely variable. In addition transitions between major associations tend to be gradational and irregular.

3.6.1 Montane forests

True lowland forest, as defined by the presence of *Dacrydium cupressinum* (rimu), is confined to the lower Cropp.

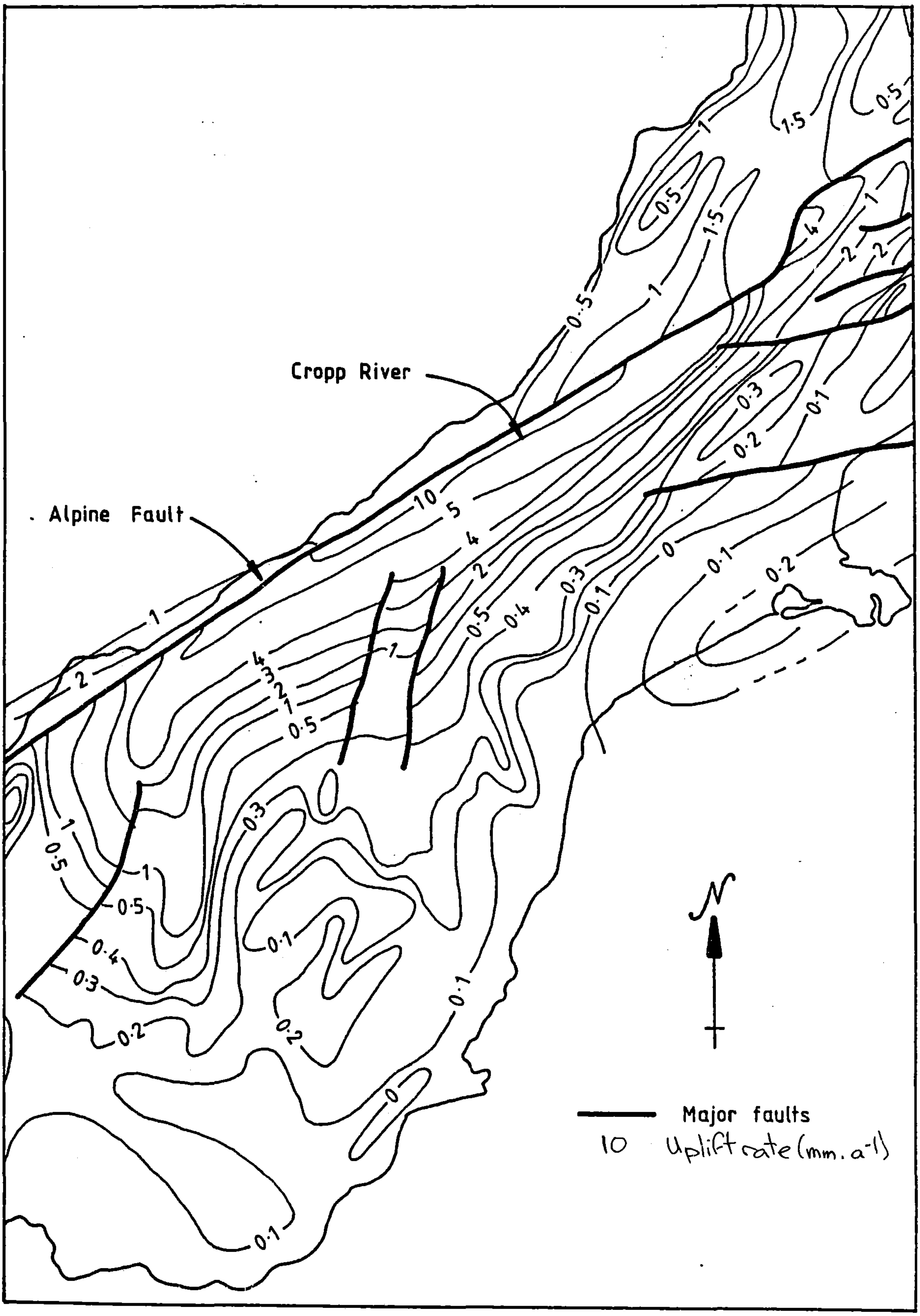


FIG. 15: Uplift map of South Island (from Wellman, 1979).

An area of tall *Podocarpus dacrydioides* (kahikatea) dominated forest occurs on the flats at the confluence of the Cropp and Whitcombe Rivers. *Libocedrus bidwillii* is present along the margins of this stand. On the hillslopes above, to an altitude of 350-400 m, a mixed conifer-hardwood forest dominated by rimu occurs. Other conifers present include *Podocarpus spicatus* (matai), *P. ferrugineus* (miro), *P. totara* (totara) and kahikatea, which are emergent over a hardwood canopy dominated by *Weinmannia racemosa* (kamahi). Forest understories are dense with a wide variety of shrubby species, tree and ground ferns.

On mid-altitude slopes (400-600 m) forests are dominated by kamahi and rata (*Metrosideros umbellata*). Scattered emergent miro occur at the lower altitudinal limits of this association while *Podocarpus hallii* (Halls totara) becomes increasingly abundant with altitude. The hardwood *Quintinia acutifolia* is abundant throughout this association and locally may achieve co-dominance with kamahi and rata. An area of relatively pure, even-aged rata occurs in the lower Cropp (NZMS 1 S64, 545113) on the site of an ancient landslide (Fig. 16).

Rata usually begins to dominate the forest above about 600 m while kamahi thins out. *P. totara* remains abundant in association with *Libocedrus*. Between 600 m and 800 m the main community recognizable, is rata - *Libocedrus* forest. Between 700 and 1000 m *Libocedrus* low forest, characterised by the absence of rata, is common. A low canopy of Compositae and Epacridaceae shrubs with emergent, often flag-form, *Libocedrus* is typical. This forest type is transitional to the subalpine scrub communities.

At all altitudinal levels throughout the forests (and scrublands) there are seral vegetation communities on unstable gullies, landslides and along stream banks. A variety of species may be present depending on altitude, site exposure, seed availability, browsing pressure, etc. Heath and mire communities also occur at all altitudes on poorly drained,

FIG. 16: Photo of lower Cropp showing even aged stand of rata signifying a rock avalanche (centre of photo). Surrounding avalanche is mid-altitude rata-kamahi forest with many dead trees. Note fluvial dissection and debris avalanches on steep (40-60°), rectilinear hillslopes.

often gently sloping sites. Stunted low growing *Dacrydium biforme* and associated shrubby species (e.g., *Olearia colensoi, Dracophyllum* spp.) and grasses (*Chionochloa rubra*) commonly occur on these sites.

3.6.2 Subalpine scrub

Forest grades irregularly into subalpine scrub at approximately 900 m. *Libocedrus* dominated stands are replaced by one of several subalpine scrub communities. Wardle (1960) recognises 5 communities. *Hoheria glabrata* low forest occurs on young, moist, freely drained, deep soils. *Olearia* low forest is found on older soils that are moist, deep and freely drained. *Olearia colensoi* scrub occupies shallow, usually immature, freely drained soils on steep slopes. *Dracophyllum uniflorum* scrub occurs on shallow soils on exposed ridges and spurs. It is commonly found at high altitudes and often merges into grassland, with the shrub component predominant on steep slopes and the grasses on gentler slopes. *Dacrydium biforme* scrub is conspicuous where soils are poorly drained. In detail many more scrub communities can be recognised.

3.6.3 Alpine grasslands

Alpine grasslands replace high altitude shrub communities at altitudes ranging from 900 m to 1400 m depending on aspect, exposure and soils. Wraight (1960) recognises 10 distinct associations characterised by the occurrence of one or more dominant species. These are: *Chionochloa rubra* on poorly drained soils; *C. flavescens* on steep slopes and young soils; *C. pallens* on steep slopes with older soils; *C. crassiuscula* at high altitudes or in late lying snow patches; *C. oreophila* on very shallow, well drained soils; *Poa cockayniana* in avalanche chutes; *Notodanthonia setifolia, Poa colensoi* and *Rostkovia gracilis* associations are considered to be induced by grazing, mainly of *C. crassiuscula* association; *Carpha alpina/Oreobulus pectinatus* association on deep, peaty, poorly drained soils.

3.7 INTRODUCED ANIMALS

Red deer (*Cervus elaphus*), chamois (*Rupicapra rupicapra*) and the Australian brush-tailed opossum (*Trichosurus vulpecula*) are the main introduced animals in the study area. Red deer were well established throughout the Hokitika River in the 1930's and probably peaked in the early 1950's when culling started (Holloway, 1966). By 1973 deer populations had been controlled by poisoning and commercial recovery to the extent that depleted vegetation in the browse zone was recovering (James *et al*, 1973). At this time Cropp River was reported as having the highest deer densities in the Whitcombe River basin - however, these densities were considerably lower than previously recorded. Deer had a marked impact on both forest understory and grassland species composition at the time that populations were high but presently the vegetation is recovering (D. Craib, N.Z.F.S., pers. comm. 1984).

Chamois have been present in the Hokitika drainage basin for at least 50 years but have never been regarded as a major problem. Numbers have generally been controlled by commerical game recovery (Pekelharing, 1973).

Opossums have been considered a major problem in Westland because of their impact on rata-kamahi and other forest types. In the Hokitika area they were liberated near Lake Kaniere in 1902 and subsequently near Hokitika Gorge in 1905 and 1912 (Pracy, 1974). They have now spread over the greater part of the entire Hokitika drainage basin although the pattern of colonisation and density is not known. Opossum numbers probably peaked in the Cropp in the early 1970's (D. Craib, pers. comm. 1984). Pekelharing (1973) surveyed the Whitcombe River and found highest opossum numbers within Cropp basin. Currently numbers are at a low to moderate level (D. Craib, pers. comm. 1984). Most of the opossum population lives within the forest, but many range upwards into the subalpine scrub and tussock zones. Their main impact appears to have been in the forests, particularly in the mid and upper forest zones (400-800 m) where understorey species composition has been modified and canopy trees have been defoliated. This zone characteristically also shows the highest deer populations, and has been regarded as the most susceptible to vegetation damage (James *et al*, 1973; Holloway, 1966).

Opossums have been cited as the primary cause of the conspicuous present day canopy mortality of rata-kamahi forests (Chavasse, 1955; Holloway, 1959, 1966; James *et al*, 1973; McCaskill, 1973; Pekelharing, 1979) although recent research indicates that natural stand dynamics may also be important (Stewart and Veblen, 1981a, b, 1982a, b; Veblen and Stewart, 1982a). In places forest deterioration is believed to have led to accelerated erosion by opening up the canopy and decreasing ground cover (Chavasse, 1955; Holloway, 1959; Pekelharing, 1979). However little substantive evidence is available to support this claim. Dieback of rata and kamahi alters species composition, but generally regeneration occurs rapidly (Allen and Rose, 1983) to maintain a dense vegetation cover.

Aerial poisoning of opossums was carried out in a large part of the Whitcombe valley in 1984 (east bank from Frews Creek to headwaters, west bank from Wilkinson to Prices River). That the Cropp River was not poisoned implies little concern over opossum numbers at this time. It should be noted that most of the detailed studies to be described were undertaken in the low forest - subalpine scrub zone that is above the zone most affected by opossums. Little rata and kamahi are present although dead standing *Libocedrus* are prominent. Based on the evidence presented in this section this part of Cropp River is regarded as little disturbed by browsing animals.

3.8 SOILS

Soils in Cropp basin were mapped in the reconnaissance survey of the South Island (N.Z. Soil Bureau, 1968a) as belonging to the Otira and McKerrow soil sets. These are described as steepland soils related to the upland and high country podzolised yellow-brown earths and podzols. Soils of both sets are formed from schist parent materials under high rainfall (1500 to 7500 mm a^{-1}). Otira soils are formed at lower elevations (150 to 1200 m) under rata-kamahi-rimu forest. McKerrow soils are formed at higher elevations (900 to 1500 m) under snow tussock grassland, subalpine scrub and herbfield. No detailed studies of the diversity of soils that occur in these sets has previously been published. Most of the soil studies in this thesis were undertaken in the low forest-subalpine scrub zone characteristic of the McKerrow set. It is considered that a similar array of soils occur in the Otira set.

CHAPTER 4 - GEOMORPHOLOGY AND EROSION

4.1 INTRODUCTION

Landforms and erosion in Cropp basin result from the interaction between tectonic uplift and global atmospheric circulation patterns in the New Zealand region. Tectonic uplift, resulting from the oblique collision between the Indian and Pacific plates, is responsible for the development of the mountain chain (3.5.4). The position and orientation of this mountain chain relative to the global atmospheric circulation pattern controls the distribution of precipitation and results in a narrow zone of extremely high annual precipitation in the western front ranges of the Southern Alps (3.3.1). High rainfall produces high rates of runoff and high rates of erosion. Viewed in the long term there appears to be a balance between high rates of uplift, and high rates of erosion (Hawkes, 1981; Griffiths and McSaveney, 1983a).

Cropp basin shows a high degree of fluvial and mass movement modification of a formerly glaciated valley. A valley-in-valley form is broadly evident with landforms of inferred glacial origin forming prominent features above the present valley floor. Glacial history is an important influence on the broad features of basin geomorphology and provides maximum ages for landforms within the basin. Glacial features are easily recognised in headwater areas, which have only recently been deglaciated, but have largely been modified in the lower reaches of Cropp River. Fluvial and mass movement processes are responsible for most present day landforms. The broad features of basin geomorphology are discussed in terms of glacial history, forms of erosion and geological controls.

4.2 GLACIAL HISTORY

4.2.1 Chronology

During the Pleistocene and Aranuian the Southern Alps were affected by numerous glacial episodes, of varying magnitude. Development of a precise chronology of glacial fluctuations has been difficult because of the paucity of finite and significant dates, the discontinuity of remnant glacial deposits, and uncertainty in correlation between valley systems of the Southern Alps.

A summary of the chronology of late Pleistocene and Aranuian glacial advances recognised in Westland is given in Table 3. This discussion is limited to the advances of the Otira Glaciation and Aranuian since the evidence of older events has been destroyed in Cropp basin. The Otira Glaciation is inferred to have begun approximately 75,000 years ago (Suggate, 1974). Three main advances are recognised (Gage and Suggate, 1958; Suggate, 1965; Suggate and Moar, 1970) named Kumara 2_1, Kumara 2_2, and Kumara 3. During all these advances valley glaciers in the Hokitika drainage basin extended beyond the mountain front as a series of ice lobes (Fig. 17).

Following the Kumara 3 advance there was rapid retreat of glaciers with periods of minor readvance. During the interval 14,000 to 8,000 years B.P. there is evidence, predominantly from the eastern Southern Alps, for several events (Chinn, 1975; Burrows, 1979; Burrows and Gellatly, 1982; Moar, 1982; Gellatly, 1984). The Waiho Loop, a prominent terminal moraine of the Franz Josef Glacier, is dated at about 12,000 years B.P. (Wardle 1973a, 1978; Mercer, 1984). Glaciers were probably restricted to the mountain valleys of the Hokitika drainage basin during these early Aranuian advances, although no moraines or depositional features have been previously identified. Impact and extent of these advances can be inferred from ice-related features and geomorphology.

There is no further direct indication of glacial advances from 8,000 years B.P. until about 4,500 years B.P. from anywhere in the Southern Alps (Burrows, 1979; Burrows and Gellatly, 1982; Moar, 1982). Since that time a series of smaller advances have occurred during a period of general glacial retreat. During these advances moraines were formed towards the heads of valleys in Westland National Park (Wardle, 1973a, 1978). While it is likely that all these advances affected Cropp basin little evidence remains to indicate their extent.

Glacial	Interglacial	Advance	Age (years B.P. except where noted)
			mid 1960s
		minor advances	early 1950s
			1930s
			1890-1920 A.D.
		several events	1600-1850 A.D.
	Aranui		1510-1095
			2570-2380
			>3980
			>4730
		probably several advances;	9000
		including Waiho Loop	12,000
		MAJOR RETREAT	
		Kumara 3_2	14,500-14,000
		3_1	17,000-16,000
Otira		Kumara 2_2	22,300-18,000
		Kumara 2_1	75,000?-35,000?
	Oturi		
Waimea		Kumara 1	
	Terangi		
Waimaunga		Hohonu	
	Waiwhero		
Porika		Porika	

TABLE 3: Summary of glacial advances recognised in Westland.

After Suggate (1965, 1968, 1974), Suggate and Moar (1970), Sara (1970), Wardle (1973a, 1978), Burrows (1978, 1979), Burrows and Gellatly (1982), Moar (1982), Mercer (1984).

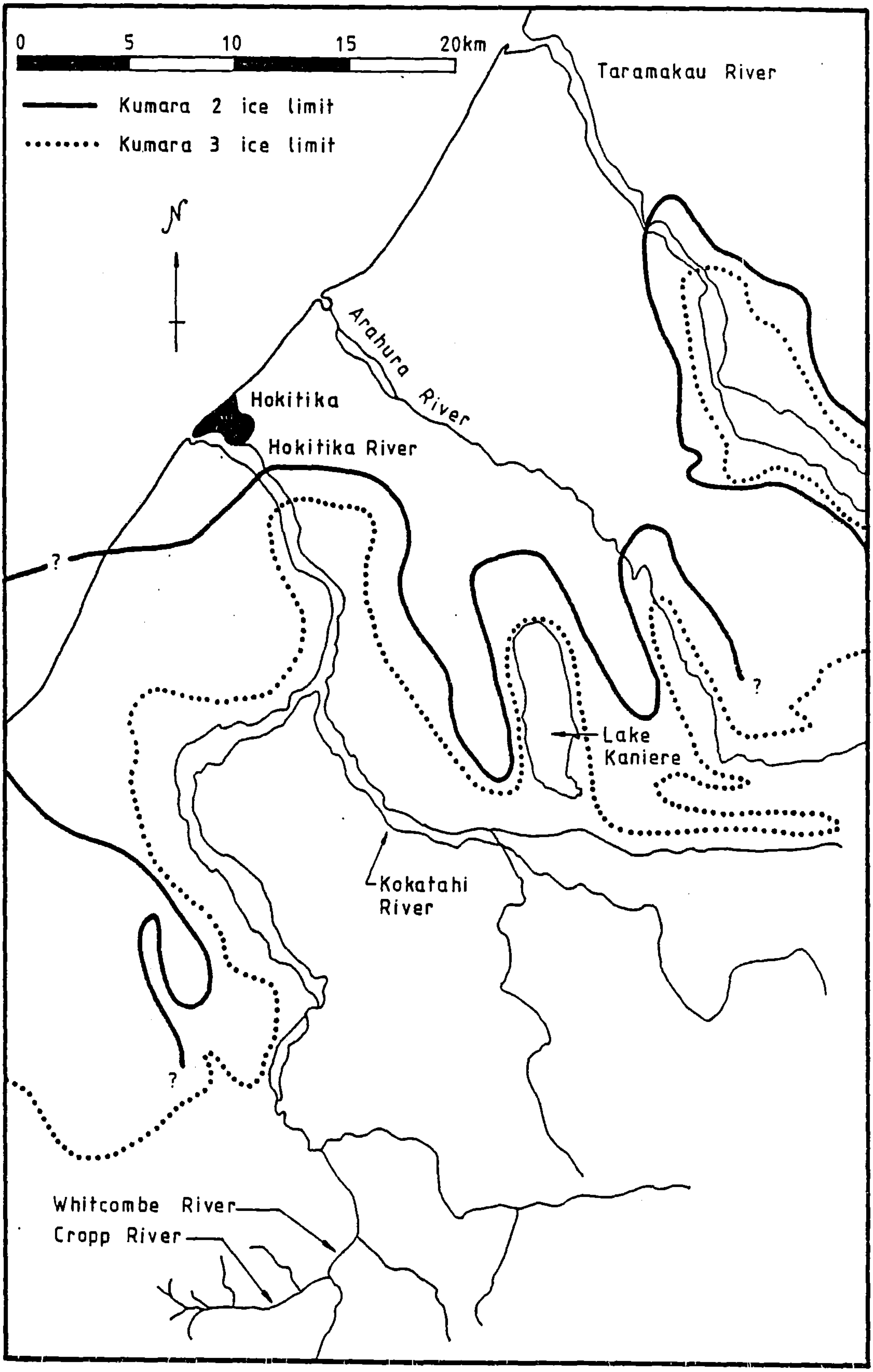


FIG. 17: Limits of ice advances of the Otira glaciation in the Hokitika area (after Suggate, 1965).

4.2.2 Effect of glaciers on geomorphology of Cropp drainage basin

In determining the effect of glacial activity on Cropp basin, in other than a general sense, a major difficulty lies in recognising former ice limits. While broad features attributable to glacial activity are recognisable (U shaped valleys, ice scoured surfaces, truncated and overriden spurs, hanging valleys) most have been modified by fluvial and mass movement erosion. Small, widely scattered remnants of former ice smoothed valley walls are the most widespread signs of former glacial activity. Few of the features that would accurately locate former ice limits (Chinn, 1979) are present. Continued rapid uplift, however, has ensured that there is significant vertical separation between landforms of the older advances. Till has been identified at two locations.

During the advances of the Otira Glaciation large valley glaciers from the Hokitika drainage basin extended to about and below present sea level (Suggate, 1965; Norris, 1978). Ice extended beyond the present coast in all except the last major advance which deposited terminal moraine near Kaniere (Fig. 17). There is no direct evidence for the age of this moraine but it is correlated with the Kumara 3 advance by Suggate (1965).

Smoothed bedrock surfaces on Steadman Brow (fig. 18), Cropp Brow and in Noisy Creek, remnants of truncated spurs (Fig. 19) and aligned ridge tops are the oldest signs of previous valley profiles in Cropp basin. These slope facets correlate with high altitude (900 m plus), glacially scoured bedrock surfaces in the Whitcombe valley and therefore are probably of glacial origin. Hawkes (1981) inferred these slope facets were formed during the last advance in which ice moved beyond the mountain front and on to the Hokitika plain (Kumara 3). Because of the extent of modification of this glacial surface no direct evidence is available to support this correlation. There is, however, some evidence to suggest that these oldest slope facets may be older than Kumara 3. Assuming that there has

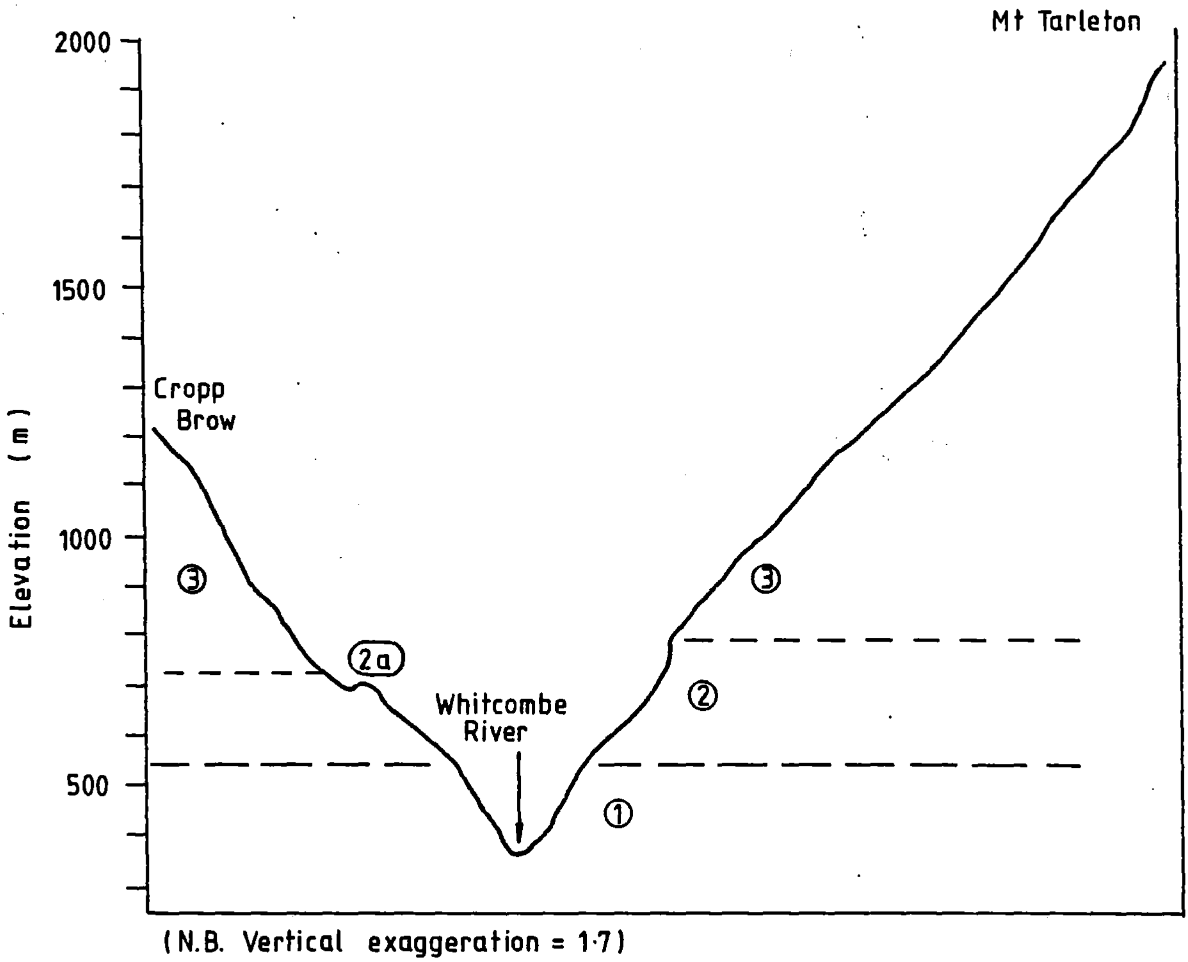
FIG. 18: Smoothed bedrock surfaces, Steadman Brow and Cropp Brow.

FIG. 19: Remnant slope facets from previous valley cutting episodes (hillslope between Hut Creek and Whinging Creek). Truncated spur with grassland vegetation (right centre of photo) is probable Kumara 2_2 correlative. Between this surface and a lower glacial bench (bottom left of photo) is the Kumara 3 surface.

been a long term uplift rate of about 10 mm a^{-1}, and that river downcutting has balanced uplift, then the valley formed during the Kumara 3 advance should be 140 to 180 metres above the present valley level. The surface identified by Hawkes is some 300-400 metres above present valley level and would therefore appear to be significantly older. Subtle evidence (breaks in slope, offset streams) suggesting the presence of ice in the Whitcombe valley below the surface interpreted by Hawkes as representing the Kumara 3 advance is present between the Price River and Brow Creek (Fig. 20). The altitude of these features (150-180 metres above present river level) is more consistent with the Kumara 3 advance. This interpretation ascribes more of the valley cutting in Cropp and Whitcombe Rivers to glacial activity than did Hawkes (1981) and suggests the oldest evidence of glacial activity in Cropp basin probably dates from the Kumara 2_2 advance.

Remnants of a U-shaped valley incised into these oldest surfaces (Fig. 19) are prominent in Cropp basin as far downstream as Reckless Torrent. While it is likely that these slope facets were formed during the Kumara 3 advance, and that ice extended to the Whitcombe valley, rapid downcutting by Cropp River has destroyed all signs of glacial activity below Reckless Torrent. Tarkus Knob, a bedrock promontory in the floor of Cropp valley at Hut Flat (Fig. 38), was probably overridden by ice during the Kumara 3 advance.

Following the Kumara 3 advance, during which ice extended to Kaniere, glacial retreat was rapid. Till (terminal moraine) dated at 10,250 $\pm$ 150 years B.P. (NZ 6576) is present in Cropp basin between Reckless Torrent and Snowy Stream (Fig. 46). This implies that ice was present in upper Cropp basin, but was probably absent below Reckless Torrent. Till, inferred to be the same age on geomorphic grounds, is also present in Hut Creek at an elevation of approximately 950-970 m i.e. at a similar elevation to Tarkus Knob (Fig. 39). An upward sequence, from kame terrace to lateral moraine to basal till (Fig. 21) indicates glacial advance. This till suggests that



KEY

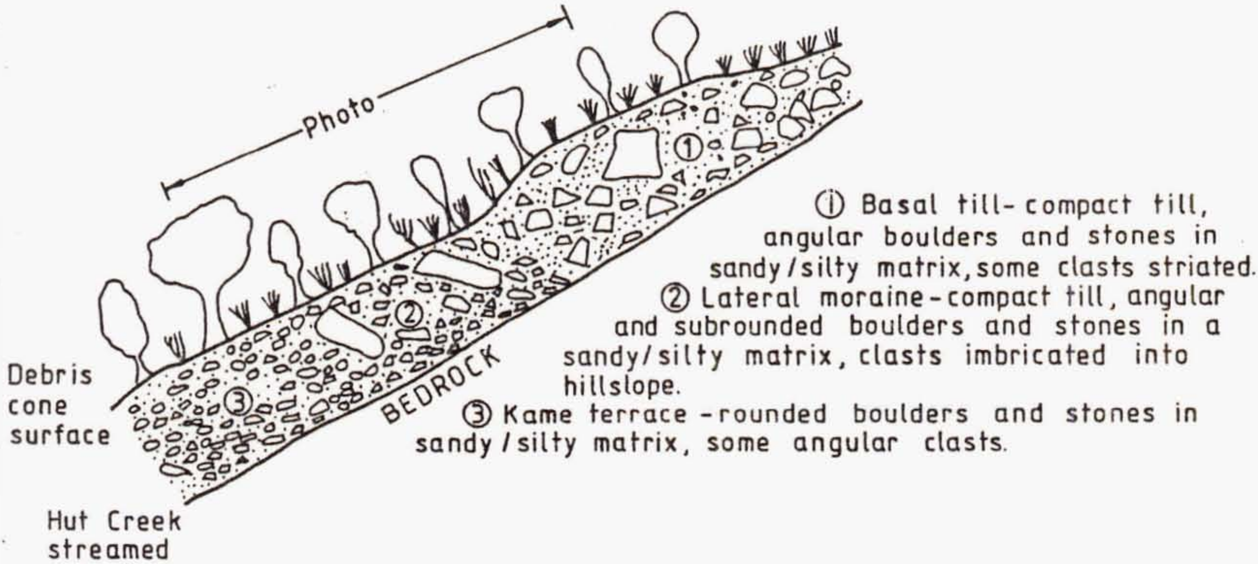
1. Post-glacial V-shaped gorge.
2. Younger, U-shaped glacial valley.

(2a) Offset tributary, possible ice limit.

3. Older, U-shaped glacial valley.

FIG. 20: Cross section of Whitcombe River from Mt Tarleton (NZMS1 S65/601077) to Cropp Brow (NZMS1 S65/551105) showing ice-related features.

a) Stratigraphic sequence.



b) Photo looking down Hut Creek - lateral moraine with clasts imbricated into hillslope; basal till appears in extreme right.

FIG. 21: Moraine sequence in Hut Creek.

during an early Aranuian glacial advance ice filled Hut Flat to about the height of Tarkus Knob (and equivalent surfaces on the south bank of Cropp River) and extended downstream to near Reckless Torrent.

Further rapid retreat followed this advance. A date from the base of peat overlying alluvium near Cropp Hut (see Chapter 5) indicates that Hut Flat was deglaciated by 7060 $\pm$ 110 years B.P. (NZ 6115), while Noisy Creek was substantially deglaciated by 7,780 $\pm$ 110 years B.P. (NZ 6348). These dates are minimum estimates for deglaciation, which could have occurred significantly earlier, but because of very limited preservation of glacial features no better estimate is possible.

Late Aranuian glacial advances were probably restricted to the cirque basins at the heads of Cropp River, Noisy Creek, Rotunda Creek and Reckless Torrent. These are only weakly modified by fluvial and mass movement processes indicating relatively recent deglaciation. The cirque at the head of Cropp River was largely deglaciated by 2,370 $\pm$ 70 years B.P. (NZ 6317), and it is inferred that other valley head cirques had a similar history.

Small glaciers are still present on the flanks of Mt Beaumont. In Rotunda Creek there is evidence for recent expansion of these glaciers (a trim line in the vegetation and a small moraine). This most probably occured within the last 100 years and has been followed by retreat.

In summary successive glacial advances have progressively deepened and widened Cropp basin to create the broad valley outline. Clearest and most widespread signs of former glacial activity are found in the upper reaches of Cropp basin where insufficient time has elapsed for fluvial and mass movement processes to modify glacial features. Older glacial features in the lower reaches of Cropp basin are highly modified, thus determination of glacial history and ice limits must be based on inference and little evidence (e.g. till). Cryonival processes (associated with frost action and snowfall; Washburn, 1973) are active today at high elevations within the basin.

4.3 FORMS OF EROSION IN CROPP BASIN

Erosion in Cropp basin includes fluvial and mass movement forms characteristic of both humid temperate and cold (alpine) climates. An attempt has been made to document the forms of erosion observed in the basin, and to qualitatively estimate their relative importance as slope modifying agents. Only mechanical forms of erosion are considered. No information is available regarding the significance of chemical denudation, It is probably high, but of minor significance relative to the extremely high rates of mechanical denudation. Evidence of active erosion is widespread throughout the basin.

4.3.1 Fluvial erosion

Fluvial erosion is the most important in basin denudation. Sheet wash occurs on bare rocky areas at high altitudes, but because of surface roughness, microchannelised flow and rilling are more common. Rills were observed on many bare rock surfaces (Fig. 22) in incompetent lithologies (e.g. pelites, greenschists, highly fractured rocks). Flow in small channels is important on vegetated slopes, with both organic and mineral material being moved (Fig. 23).

These small features contrast with the larger gullies and ephemeral streams. The high degree of fluvial dissection of the basin (Fig. 24), with many ravines cut in bedrock (Fig. 25), reflects the importance of fluvial erosion in the development of this landscape. To some extent the degree of dissection is masked by the dense plant cover of forest and grassland, over much of the basin (Fig. 26). Drainage density based on all ephemeral channels would be at least an order of magnitude greater than the measured value of 5 $km.km^{-2}$, but obtaining the data for the calculation is impossible at present (lack of suitable topographic base map).

a) Beaumont Basin

b) Danger Gully

FIG. 22: Rilling in pelitic schist.

FIG. 23: Slope wash amongst tussock grassland.

FIG. 24: Intense fluvial dissection of pelitic schist, Danger Gully. Hut Flat and Tarkus Knob in right foreground-location of S8 and S9 indicated by arrow.

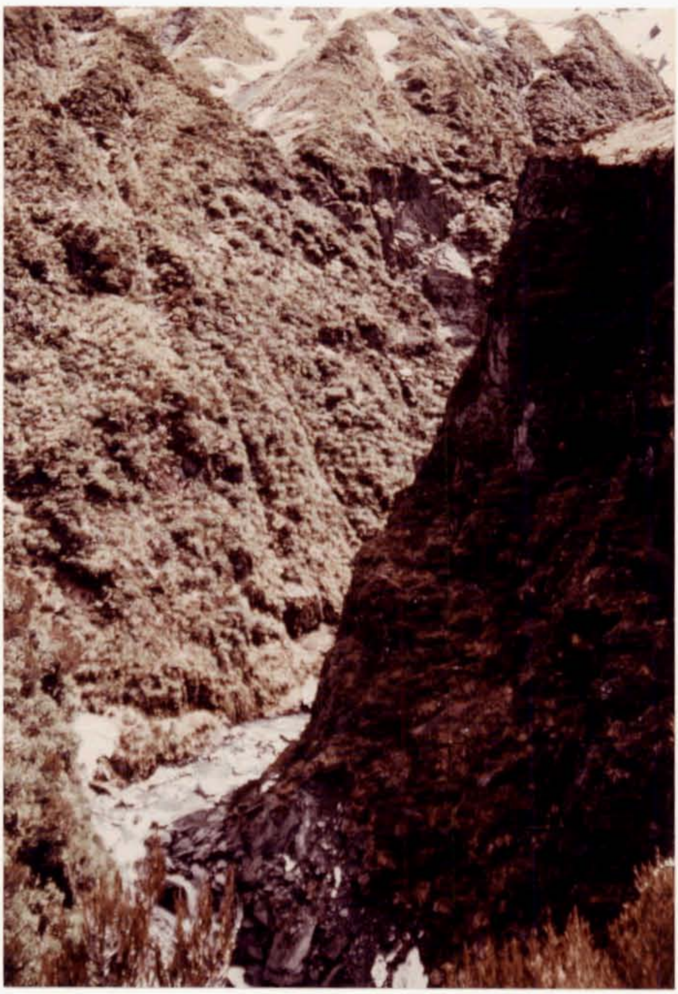


FIG. 25: Ravine cut in bedrock, Reckless Torrent.

a) Tussock grassland (author standing in collapsed pipe)

b) Scrub-covered area (with sunlight angle enhancing dissection)

FIG. 26: Vegetation masking fluvial dissection

Streambank erosion is important in Cropp basin. Because of the many stream channels small areas of streambank erosion are common and, in combination with active mass movement, provide for rapid delivery of sediment derived from hillslope erosion into the fluvial system.

Subsurface pipes in soils are also common. These range from a few millimetres to 30 centimetres in diameter and some have collapsed to form deep, steep sided channels (Fig. 26a).

4.3.2 Mass movement erosion

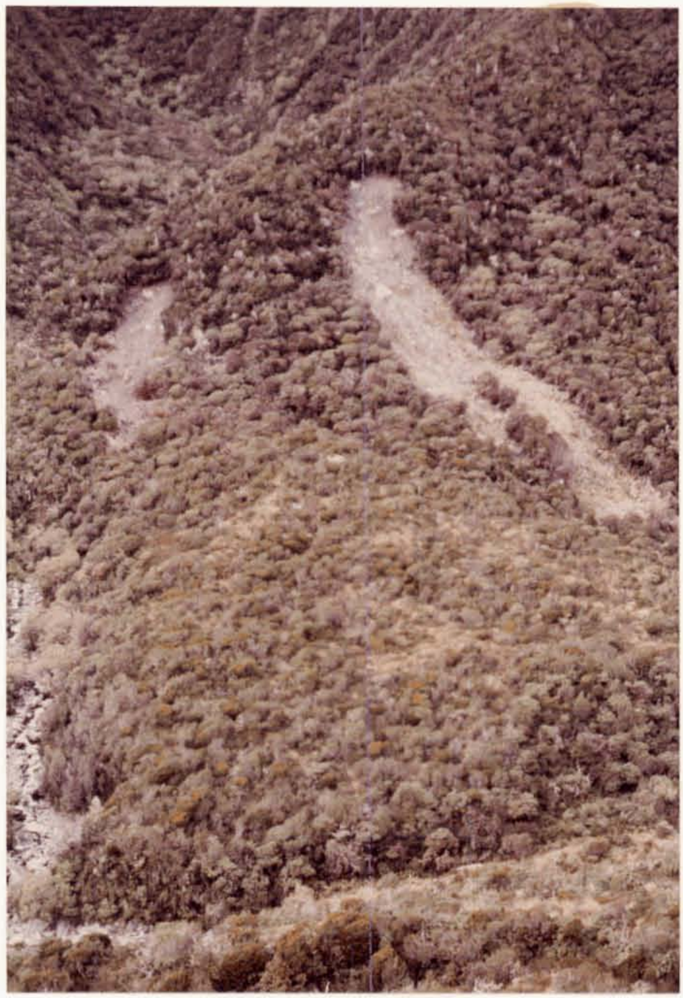
Mass movements in Cropp basin involve both bedrock and regolith. Rockfall is common in any areas where bedrock crops out, but particularly at high elevation. It includes simple rockfalls, confined to the removal of individual, surficial blocks from cliff faces, and slab and toppling failures associated with steeply dipping schist (Fig. 27) and the Pounamu ultramafics (Fig. 28). A small rock avalanche in the lower Cropp (Fig. 16) is dated at younger than 363 $\pm$ 56 years B.P. (NZ 5252) and involved approximately 2 x 10^5 m^3 of material. Rock slides and slumps in Cropp basin are subtle features but are more obvious in the Whitcombe valley (Fig. 29). The failures in the Whitcombe are large features that involve entire hillsides and may have been intermittently active for thousands of years. It is likely that bedrock failures were most important in Cropp basin immediately following glacial retreat when removal of lateral support would have been followed by stress release and opening of joints and fractures. Debris cones (see 5.2) on the south side of Cropp River were probably largely constructed at this time. Similar smaller debris cones have formed in the cirque at the head of Cropp River following more recent deglaciation.

Regolith failures are dominantly episodic flows and slides, associated with high-intensity rainfalls. In the forest and scrub zones, these most commonly are debris avalanches (Fig. 30a, b) although debris flows and slides also occur. Debris avalanches observed during the period of this study (9 in

FIG. 27: Slab failure and rockfall associated with vertical schistosity, Top Cropp.

FIG. 28: Slab and toppling failures, Pounamu ultramafics, near Noisy Creek.

FIG. 29: Rock slide, adjacent Collier Gorge, Whitcombe River.

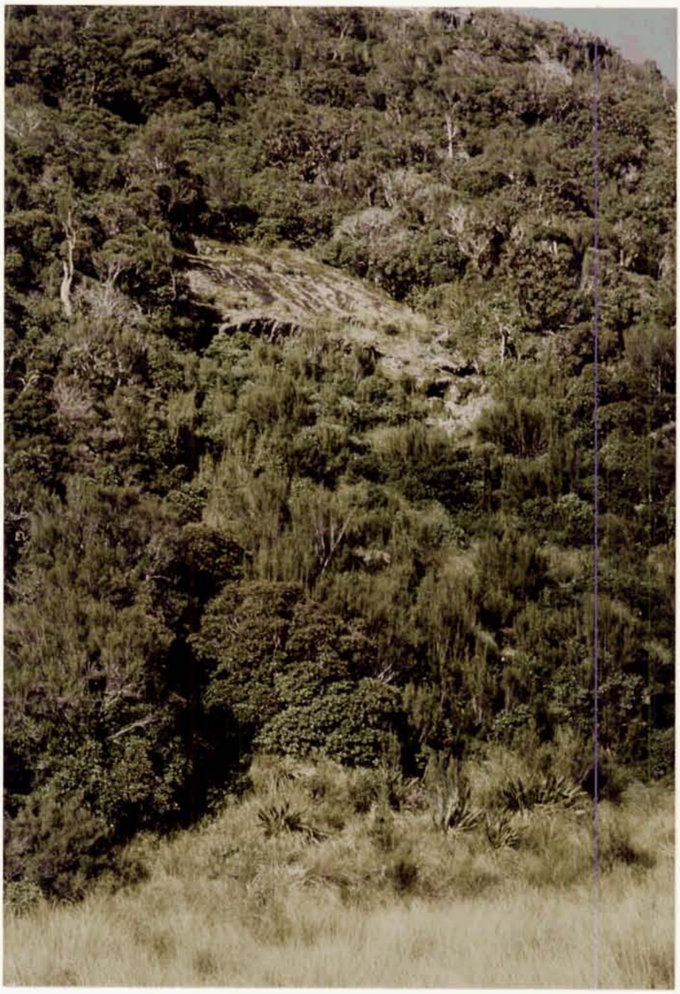


a) Debris avalanches, adjacent Whinging Creek.

b) Debris avalanche with rilling, lower Cropp.

FIG. 30: Mass movements of regolith.

c) Debris slide, Tarkus Knob.



d) Debris flow, Noisy Creek (flow has come down creek and across middle of debris cone).

total) commonly involved of the order of 10^2 m^3. One older debris avalanche (c. 140 years) behind Cropp Hut (Fig. 39) involved an estimated 1-2 x 10^4 m^3. Debris slides (Fig. 30c) occurred where there was a distinct contact between bedrock and overlying regolith. Most failures probably involved both sliding (particularly near the crown of the failure) and flow and would be best described as debris slides-avalanches. Location of the crown of most failures appeared to be controlled by proximity of bedrock to the ground surface, resulting in localised increase in pore water pressure and loss of shear strength. Evidence of debris flows was observed in Noisy Creek (Fig. 30d) and Hut Creek. At both locations they probably occurred where there was rockfall into steep streams during storms. They were characterised by the presence of large (up to 5 metre diameter) blocks of bedrock at the terminus and the development of levees at the margins of the flow. Griffiths and McSaveney (1983a) suggest many of the tributary streams owe their channel form to debris flows.

In contrast to the long, narrow debris avalanches of the forest and scrub zones, mass movements in the grassland areas of the basin were mainly small scale, localised debris slides-avalanches. As the numbers of such movements are very large, these small scale features may be as important as the more prominent debris avalanches in terms of overall basin denudation.

Slow, continuous mass movement erosion (soil creep and solifluction) is also important. The most conspicuous signs of this type of erosion were turf banked terraces (Davies, 1972) in the tussock grasslands (Fig. 31). How these features develop, and whether they are currently active, was not investigated in detail. They are, however, very common features throughout the tussock grasslands, on both steep and gentle slopes (observed on slopes of 5° to approximately 30°). Movement involves flow of saturated soils, but freeze-thaw may also be important. Tension cracks across steep slopes, and particularly where changes in slope from lower to higher slope

a) Surface form, 10° slope

b) Section through turf bank showing topsoil buried under lobe.

FIG. 31: Turf banked terraces, Noisy Creek.

angles occur, also indicate creep. Hawkes (1981) identified creeping failures affecting surficial rock debris in Danger Gully.

Snow avalanches, slush flow, and related snow movement erosion (Davies, 1972) are other forms of mass movement common in Cropp basin. These include small, localised movements and large ridge crest/valley bottom movements. Dirty snow avalanches occur on steep slopes in late spring/early summer and incorporate varying amounts of organic and rock debris. Snow sliding or creeping in gullies also contributes to downslope transport of rock debris (Fig. 32).

Tree windthrow also contributes to downslope movement of soils. Windthrown trees are scattered throughout the forest and scrub covered areas of the basin. This form of erosion generally affects small areas and disturbs the upper 20-50 cm of soil.

4.3.3 Aeolian erosion

Evidence of aeolian erosion in Cropp basin is limited to windblown schist fragments in the tussock grasslands. On the saddle at the head of Noisy Creek platy fragments of schist up to 2 kg in weight were observed lying on living vegetation. There are no streams in the area, nor cliffs from which the material could have fallen. It appeared that the fragments had blown uphill from a steep cliff facing directly into the northwest wind, attesting to the power of the wind. Wind erosion is probably a minor form of erosion in this environment but is capable of detaching and transporting material in suitable locations.

4.4 GEOMORPHOLOGY OF CROPP BASIN

4.4.1 Geomorphic regions

Landforms in Cropp basin reflect glacial history, extensive post-glacial modification, and lithologic and structural controls. The main features are U-shaped cirques

FIG. 32: Snow creep and avalanche debris, Reckless Torrent.

at the heads of larger tributaries, and V-shaped valleys with uneven profiles in the lower parts of tributaries and the main valley. Most landforms are erosional features although significant deposition has occurred locally. Five major regions can be recognised (Fig. 33): lower Cropp (with dominant fluvial downcutting), mid-Cropp (with significant depositional landforms), upper Cropp (tributaries with remnant glacial surfaces), pelitic schist region, alpine region.

(a) Lower Cropp region (Fig. 34).

In its lower reaches (below Reckless Torrent) Cropp River flows through a narrow V-shaped gorge incised more than 500 metres below the surrounding ridges. Riverbed gradient averages 0.13, but is very uneven (Fig. 35). Gradient increases downstream and is steepest in the bedrock-floored section between Pounamu Creek and where the Cropp flows into the flat-floored Whitcombe valley (averaging nearly 0.20). The river flows in a channel composed dominantly of boulders, with diameters commonly in excess of 1 metre (and up to about 10 m). Size of bed material partly reflects a lag formed of the coarsest sediment delivered by steep tributaries.

Hillslopes are long (500-1000 m), steep (generally in excess of 40°) and rectilinear. This region is dominated by fluvial downcutting with a parallel drainage network on hillslopes. Three major gullies on the north side of Cropp River carry ephemeral streams, draining from an area of greenschist undergoing rapid erosion. Fluvial dissection and debris avalanching, as a result of constant undercutting of steep slopes, are the main forms of slope erosion (Fig 16). Prominent unvegetated scars, and older scars covered in seral vegetation indicate frequent mass movement. The dated rock avalanche (4.3.2) is located in this region.

There are no unequivocal signs of former glacial activity in this region. Smoothed ridge tops on Cropp Brow (Fig.

FIG. 33: Geomorphic regions of Cropp basin.
New Zealand Aerial Mapping Survey No. 1580, Run No. 3723/14. Date 19/3/65. Approximate scale 1:34,000.

1) Lower Cropp region.
2) Mid-Cropp region.
3) Upper Cropp cirque basins.
4) Pelitic schist region.
5) Alpine region.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NOTE - long, steep, rectilinear hillslopes
- steep riverbed gradient
- large boulders forming channel
- even aged stand of rata (relating to rock avalanche)

FIG. 34: Lower Cropp region.

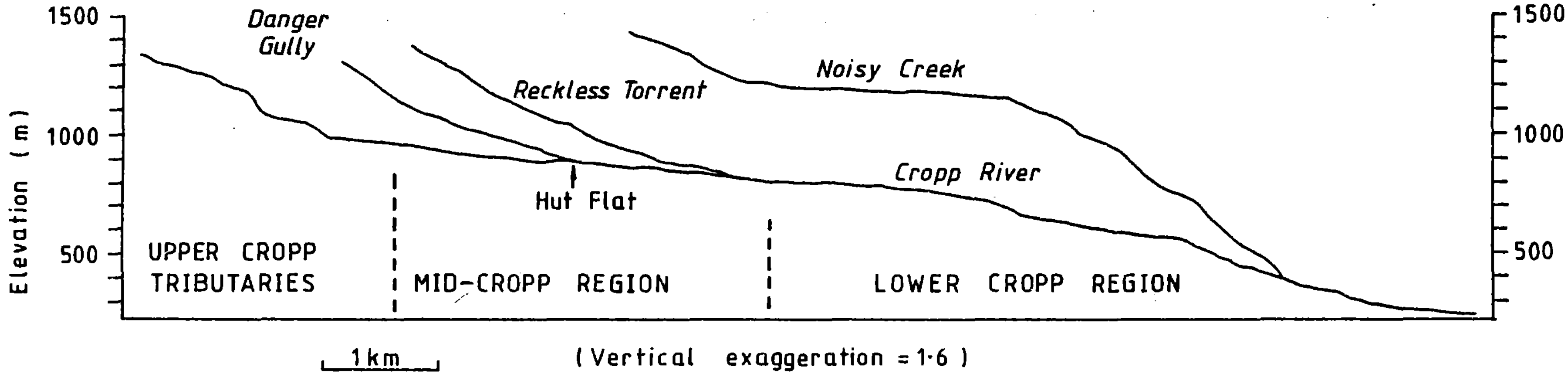


FIG. 35: Longitudinal profile of Cropp River.

18) may be the result of glacial activity although no bare rock surfaces with glacial striations were observed. Additionally no clearly old soils were found on this ridge, suggesting that if these surfaces had a glacial origin they have been modified by post-glacial erosion.

This part of the basin has been deglaciated the longest, however it also has been most affected by base level change caused by downcutting in the Whitcombe valley.

(b) Mid-Cropp region (Figs. 38, 39, 46, 47).
Cropp River is characterised by a more gentle stream gradient, a wider valley and significant depositional features in the reach between Reckless Torrent and Rotunda Creek (mainly on the south bank), and in the lower reaches of Rotunda Creek. Average stream gradient in this region is .05 and is relatively uniform. The river flows through a series of flats and rapids between Reckless Torrent and Snowy Stream and is incised 50 to 100 m into debris cones. Above Snowy Stream the river flows through a bedrock-floored gorge where it is confined by the resistant bedrock of Tarkus Knob on the north bank, and by large debris cones on the south bank. Between Tarkus Knob and Danger Gully the river flows across an alluvial flat (Hut Flat) with a gentle gradient (.044) before winding through alluvial cones (see 5.2) built by Danger Gully, Whinging Creek and Rotunda Creek.

Signs of former glacial activity are present throughout this region. These include the broad U-shaped valley form, smoothed valley walls and bedrock surfaces, and till (Figs. 38, 39, 46). However, many of these features are highly dissected and difficult to interpret.

Throughout this zone significant depositional features have formed as a response to more gentle stream gradient and local conditions. Large debris cones have formed on the south bank of Cropp River between Cream and Hut

Creeks, and in Rotunda Creek. Smaller alluvial terraces and cones have formed upstream of Tarkus Knob. This region was selected for most detailed study (Chapter 5) because of its array of landforms and stratigraphic record. The surrounding steep hillslopes that were the source areas for the depositional features are also included in this region. The form of the large debris cones suggests that rockfall was an important process in their formation (especially immediately after deglaciation). Now debris avalanches, debris flows, fluvial dissection and snow avalanching are probably more important forms of erosion in the source areas which have a high density of ephemeral stream channels.

(b) Upper Cropp region (figs. 48, 53).
This includes the headwaters of Cropp River, Reckless Torrent, Noisy Creek, Rotunda Creek and Beaumont Basin. These areas were occupied by glaciers during the Aranuian and U-shaped valleys are weakly modified because of the short time since deglaciation, or because of lithologic barriers to valley incision (as at Noisy Creek).

River gradients are steep (except for Noisy Creek) and extremely uneven because resistant lithologic units form waterfalls and gorges. Average stream gradient between Rotunda and the head of the Cropp is 0.18, due mainly to a steep, rock floored gorge with a series of waterfalls at the edge of the top Cropp cirque.

Deglaciation of these cirques ranges in age from very recent (Beaumont Basin) to early Aranuian (Noisy Creek) and degree of modification varies accordingly. Hillslopes tend to be extremely steep (40-90°) but significant depositional features occur in footslope positions. Rockfall, debris avalanches, snow avalanches and fluvial dissection are the main forms of erosion. Both Noisy Creek and the Top Cropp cirque were investigated and results are included in Chapter 5.

(d) Pelitic schist region (Fig. 24).
This area is underlain by highly fissile, often shattered, pelitic schist. The area on the north bank of Cropp River between Reckless Torrent and the head of the Cropp is characterised by an extremely high drainage density and short, steep, rectilinear slopes. Slope angles average 40-50° and are commonly greater towards the heads of small streams. This area is an extreme example of Cotton's (1958) fine textured (feral) erosional relief. Erosion is dominated by fluvial dissection (streambank erosion and channel incision), small flows and slides during snowmelt and high-intensity rainfall, and probably soil creep. Erosion is extremely active as a result of constant undercutting of slopes. Soils characteristic of this terrain are described in Chapter 5.

(e) Alpine region.
This is the area above vegetation limit and subject to cryonival processes (i.e. those associated with freeze/thaw and snow activity). It includes the high ridges around the basin perimeter which range from glacially smoothed broad ridge tops to steep, dissected aretes. Rockfall, freeze-thaw and associated snow patch erosion (Davies, 1972), snow avalanches, and fluvial erosion are important in modifying these areas.

4.4.2 Geological controls on geomorphology (summarised from Hawkes, 1981).

Lithology and structure in the underlying bedrock have an important influence on basin geomorphology and erosional characteristics. Resistant rock units (chert, ultramafics, thickly bedded psammitic schist) are responsible for the development of many waterfalls and gorges. Most notable in this respect are the Pounamu ultramafics which have provided a barrier to valley deepening in Noisy Creek. Resistant beds at the edge of Top Cropp cirque and Beaumont Basin have also prevented valley deepening. Areas underlain by less resistant lithologies (e.g. pelites, greenschist) are preferentially eroded and tend to be extremely dissected.

A prominent ENE-WSW striking fault set controls streams draining Galena Ridge while a second set of N-S trending faults controls drainage of Cropp Brow and in the lower Cropp gorge (Figs. 13 and 14). Amount and attitude of jointing and fracturing in the bedrock, and orientation in relation to surface topography, are important in controlling relative erodibility and hence geomorphology. A south-west dipping joint set on the north side of Cropp basin controls south-facing slope angles (Fig. 14). Structural control of weathering characteristics of bedrock is clearly evident on aerial photographs, particularly in alpine areas devoid of vegetation.

CHAPTER 5 - GEOMORPHOLOGY AND SOILS OF DETAILED STUDY AREAS

5.1 INTRODUCTION

After a reconnaissance of Cropp basin four areas were chosen for detailed study. The mid-Cropp region showed great potential for a study of soil-landform relationships and of soil development spanning a relatively long period. The area upvalley of Tarkus Knob (Hut Flat) was studied in most detail because of its array of landforms and stratigraphic record. The lower mid-Cropp region, between Snowy Stream and Reckless Torrent, also contained significant depositional landforms with dateable material. Two cirque basins in the upper Cropp region were also studied. The cirque at the head of Cropp River (Top Cropp area) provided an example of a younger landscape, deglaciated more recently than the mid-Cropp region. Noisy Creek is a tributary basin where downcutting has been retarded by resistant rock (Pounamu ultramafics) at the basin outlet. Geomorphology and Hawkes (1981) interpretation suggested this would be an area of relative stability with a useful depositional record.

5.2 METHODS

Soil development sequences were examined at Hut Flat on both gently and steeply sloping topography. Ages of most land surfaces and associated soils were estimated by dendrochronology and ^{14}C dating. Dendrochronology was used to date the onset of succession as a result of denudation of hillslopes or inundation of valley floor areas. Isolated dates provided a chronology for the time of origin of hillslopes and river terraces. Only young events (<200 years) were dated by this means. The stands dated were clearly successional and the shrubs were first generation plants, which were assumed to have begun growing within a few (<5) years of site denudation. The species dated are believed to have annual growth rings (Wardle, 1963) and all measurements were replicated. ^{14}C dating provided the basis for estimates of age of older land surfaces and associated soils. 24 dates on wood and peat samples, covering the age range from modern (<250 years) to approximately 10,000 years B.P., were obtained and used to infer a history of landform evolution, soil development and erosion. All ^{14}C dates for the Cropp study were supplied by the Radiocarbon Dating Laboratory, Institute of Nuclear Sciences, Lower Hutt, New Zealand. Dates supplied were:

- "conventional" radiocarbon age
- results calculated using new half-life
- results calculated using new half-life and applying, where possible, a secular correction (Michael and Ralph, 1972).

All dates are discussed in terms of "conventional" radiocarbon age.

Soil stratigraphic principles were used to investigate landscape history, although it was not possible to formally map geomorphic surfaces. In Cropp basin physiographic units (simply referred to as landforms) can be recognised, but they are generally of limited spatial extent or temporal persistence, or they include land surfaces of more than one age. By comparison with the soils of directly dated surfaces, soil morphology provides an indication of the chronological array of soils present on various landforms.

The main aim of the soil studies was to establish the broad soil-landform relationships and to interpret the soil pattern in relation to spatial and temporal controls. Therefore a free survey technique was adopted. The difficult terrain and dense vegetation also necessitated this approach. At Hut Flat, observations were made mainly along transects across the major landforms. In grassland areas, transects were generally along the slope contour, but in areas of scrub they paralleled the slope profile to aid accurate location of observation points on photos. At other study areas sampling was less regular and comprehensive. Transects covered representative areas of the landscape, with the density of observations dependent upon landform and associated soil changes. Soils were examined in pits, except where they were fine textured or organic, where a Dutch auger was used.

Topography is described in terms of the landform, and the shape of the slope profile and contour within each landform. The types of landform recognised in the site descriptions are:

(a) hillslopes - moderately to steeply sloping (>c. 20°) elements of the landscape with a thin (<3 m) covering of colluvium over bedrock.

(b) debris cones - generally steeply sloping (up to c. 45°), wedge shaped (in profile) accumulations of poorly sorted colluvium in footslope positions.

(c) alluvial cones - less steeply sloping cones (<c. 20°) with debris that shows significant fluvial sorting.

(d) alluvial terraces - gently sloping (<5°) linear benches parallel to river gradient. Debris is relatively well sorted.

(e) risers - short, steep slopes separating successive step-like landforms (e.g. terraces, alluvial and debris cones).

The first three terms are modifications of terms used to describe debris accumulations on hillslopes by Rapp (1960), Church *et al* (1979) and White (1981). These authors emphasised the processes involved in producing different forms (rockfall, snow avalanche activity, fluvial activity) in Arctic and alpine environments. While similar forms can be recognised in Cropp basin, the processes involved in producing them may be different (possible greater importance of fluvial activity). Thus the terms used for the description of landforms in Cropp basin are primarily descriptive rather than genetic. Shape and slope of the deposit, in conjunction with relative sorting of sediment, were used as definitive characteristics. It is recognised that the terms describe members of a continuum.

The description of slope components of landforms follows that of Hack and Goodlet (1960), Ruhe and Walker (1968) and Huggett (1975). The slope contour is subdivided, according to contour curvature, into hollows (with concave contour), sideslopes (linear contour) and noses (with convex contour). The slope profile is subdivided, according to profile curvature into:

(a) summit - uppermost portion of slope with planar to slightly convex profile;

(b) shoulder - upper portion of slope with convex profile;

(c) backslope - middle portion of slope with planar profile;

(d) footslope - lower portion of slope with concave profile;

(e) toeslope - lowermost portion of slope with slightly concave to planar profile.

Many slopes in Cropp basin are rectilinear with only the backslope element present. Location within a slope profile in these situations is indicated by the use of the terms upper, mid or lower backslope. The above terminology is not applied to flat to gentle slopes where slope profile and contour are simply described as convex, concave or linear.

Brief descriptions of soil site and profile morphology were recorded for approximately 300 locations throughout the basin. Only 19 profiles were described and sampled in detail. These soils were chosen because: (a) their ages could be estimated from stratigraphy, dendrochronology or radiocarbon dating; and (b) for practical considerations (fewer problems with water). The sampled soils are characteristic of the better drained sites in the landscape and demonstrate the maximum expression of soil development (podzolisation). Soils of wetter sites are under-represented, as are those formed directly from bedrock. Most of the sampled soils are formed from alluvium or colluvium.

The soil horizon nomenclature used is defined in Hodgson (1976) and Avery (1980) and is summarised in Appendix 2. The horizon subscript j is used to denote weak expression of a qualifying suffix (Canada Soil Survey Committee, 1978). Moist soil colours were recorded using a Munsell soil colour chart. Terms for field estimation of fine earth (<2 mm) texture are those of Taylor and Pohlen (1962). Visual estimates of the size of rock fragments (>2 mm) was made using the following classes:

gravels	2 - 20 mm (largest dimension)
stones	20 - 200 mm
boulders	>200 mm

Estimates of the volume of rock fragments followed Hodgson (1976):

slightly stony (gravelly)	< 15% rock fragments
moderately stony (gravelly)	15 - 35%
very stony (gravelly)	35 - 70%
extremely stony (gravelly)	> 70%

A subjective assessment of clast weathering was made to indicate the range of weathering grades present in each horizon. It followed the classification of Fookes *et al* (1971) - Table 4. Terms used to describe consistence, structure, mottles, roots and horizon boundaries are defined in Hodgson (1976).

Soil profile form is used as a complementary concept to that of horizonation to give a synoptic view of the textural, structural and redox properties of whole soil profiles. Cutler's (1980, 1981, 1983) terminology for profile forms is used (Fig. 36).

(a) Soil texture profile forms - Cutler (1981) defined a series of texture profile forms based on the trend of fine earth texture with depth down to, and including, the C horizon. Tonkin (1984) added a series of comparable terms to describe the depth trend of the skeletal (>2 mm) fraction.

(b) Structure profile forms - these are defined on the relative proportion and dominant structural type of peds in the A (and/or E) when contrasted with the B horizon. Soil horizons are divided into subpedal (apedal or very weakly developed peds) and pedal (strong, moderate or weakly developed peds) classes. Variation in pedality with depth forms the basis of the six recognised structure profile forms.

(c) Redox profile forms - these are defined in terms of the

Degree of decomposition of rock fragments	Grade	Description: Hard rock fragments within the debris mantle regolith. Gravel, stone and boulder size grades.
Unweathered or Fresh	1	Rock fragments show no surface discolouration, loss of strength or any other weathering effects.
Slightly weathered	2	Rock fragments discoloured on the surface, with slight penetration along defect surfaces and of the rock fabric. This may form a recognisable rind, of measurable thickness. The interior of rock fragments is the colour of fresh rock. Rock fragments are only slightly weaker than the fresh rock.
Moderately weathered	3	Rock fragments are discoloured throughout, although a core of fresh rock may persist. Rock fragments cannot be crushed by hand but may break along defects.
Highly weathered	4	Rock fragments are discoloured throughout with colours comparable with the fine earth matrix. They can be crushed by hand, yielding sandy or silty textures, with some smaller rock fragments. Some mineral components and zones of decomposition along defects may be clayey in texture.
Completely weathered	5	Rock fragments have a colour and consistence comparable to the fine earth matrix. They are recognisable only by the persistence of a rock fabric.
Soil	6	Rock fragments are indistinguishable from the fine earth matrix. The fine earth matrix may vary in texture from sandy to clayey and has a soil fabric and/or structure.

TABLE 4: Classification of weathering grades for hard rock fragments in a debris mantle regolith-modified and adapted from Fookes *et al* (1971) by Tonkin (1984).

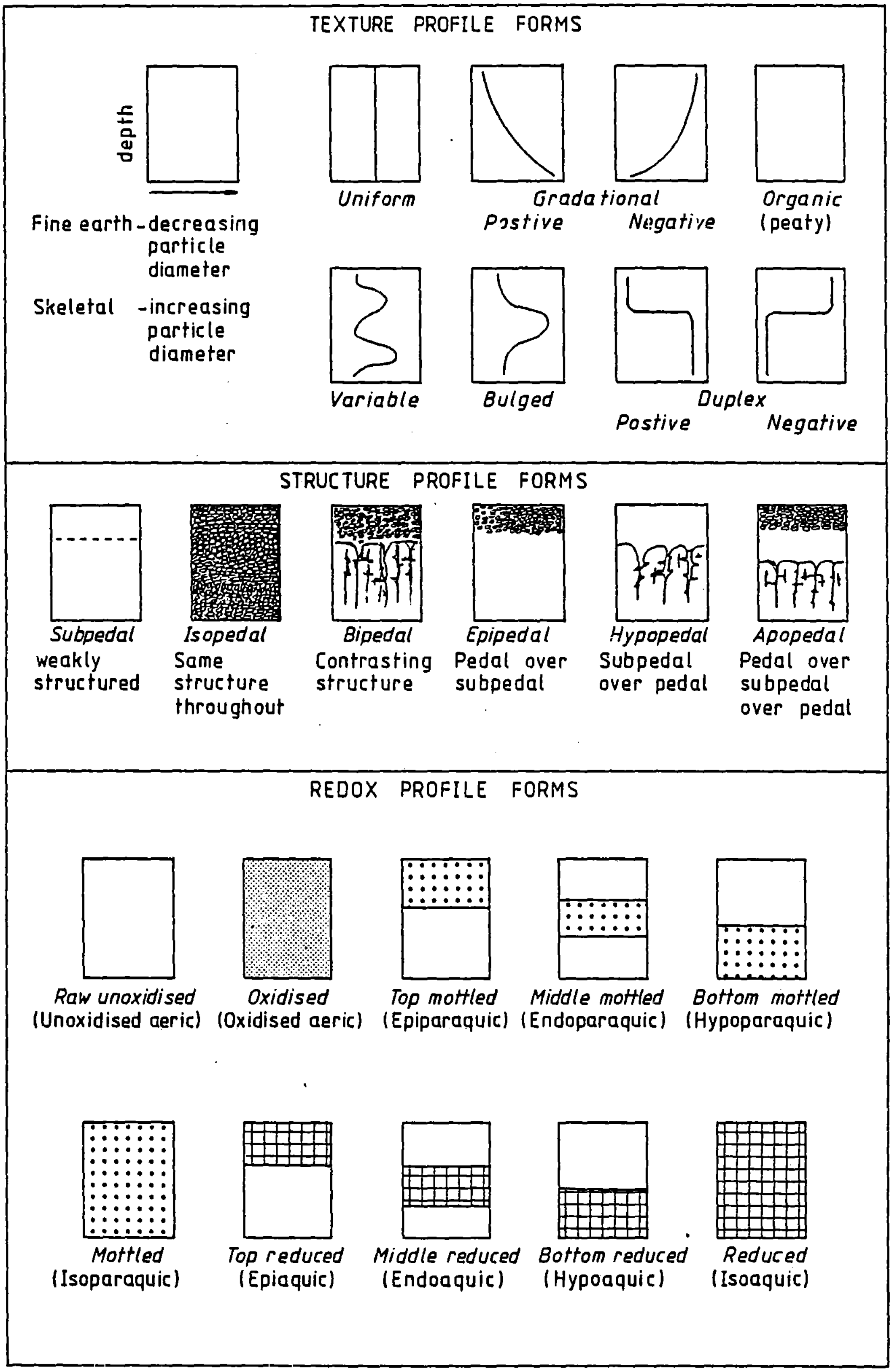


Fig. 36: Texture, structure and redox profile forms (Cutler, 1980, 1981, 1983).

dominant colour and colour patterns in the A, B and C horizons, which reflects the redox status of the soil profile. Cutler (1983) introduced a series of technical terms to describe redox profile forms. For simplicity, the common terms of Cutler (1980) are used although Fig. 36 includes both terminologies.

Throughout the discussion the common names of the New Zealand Genetic Soil Classification (NZ Soil Bureau, 1968b) are used to describe the major morphological features of observed soil profiles. The broad classes of soils recognised are recent soils, yellow-brown earths, podzolised (and gleyed) yellow-brown earths, gley podzols, organic soils, gley and gley recent soils. These classes provide synoptic terms for describing characteristic sequences of horizonation and associated characteristics. A more comprehensive account of soil classification is included in Chapter 8. Buried soils are described using the terms introduced in Chapter 2 (i.e. composite or compound).

5.3 HUT FLAT

At Hut Flat (Figs. 37, 38, 39) a series of alluvial terraces, alluvial and debris cones are preserved upstream of Tarkus Knob, where the combination of resistant bedrock and delivery of coarse (1 to 10 m diameter) sediment from Hut Creek has slowed downcutting by Cropp River. The surrounding steep hillslopes have been sources of sediment for the depositional landforms, and their pedological characteristics reflect erosion history. Fig. 40 summarises the field relationships between many of the landforms, soils and ^{14}C dates at Hut Flat. Landforms and soils of the depositional surfaces in the valley bottom are described first, followed by landforms and soils of the steep hillslopes.

A complete listing of growth ring counts and ^{14}C dates is given in Tables 5 and 6. The stratigraphy associated with each ^{14}C date is given in Appendix 1. Complete site and soil profile descriptions, and photographs of sampled soils are

Fig. 37: Vertical aerial photo of Hut Flat showing localities mentioned in the text, soil and radiocarbon sampling sites. New Zealand Aerial Mapping Survey No. C8383, Run B/5. Date 24/3/84. Approximate scale 1:3,750.

Radiocarbon dating sites.

1. NZ 6318 (<250 years B.P.), NZ 6319 (1045 ± 60 years B.P.)
2. NZ 6347 (<250 years B.P.)
3. NZ 6600 (<250 years B.P.)
4. NZ 5249 (291 ± 46 years B.P.), NZ 5250 (1080 ± 60 years B.P.), NZ 5365 (1285 ± 70 years B.P.)
5. NZ 6588 (1275 ± 45 years B.P.)
6. NZ 6113 (1700 ± 65 years B.P.), NZ 6114 (3260 ± 70 years B.P.)
7. NZ 5367 (1750 ± 45 years B.P.)
8. NZ 6341 (2470 ± 60 years B.P.)
9. NZ 6115 (7060 ± 110 years B.P.)
10. NZ 5369 (7380 ± 90 years B.P.), NZ 5368 (8390 ± 90 years B.P.)

Soil sampling sites.

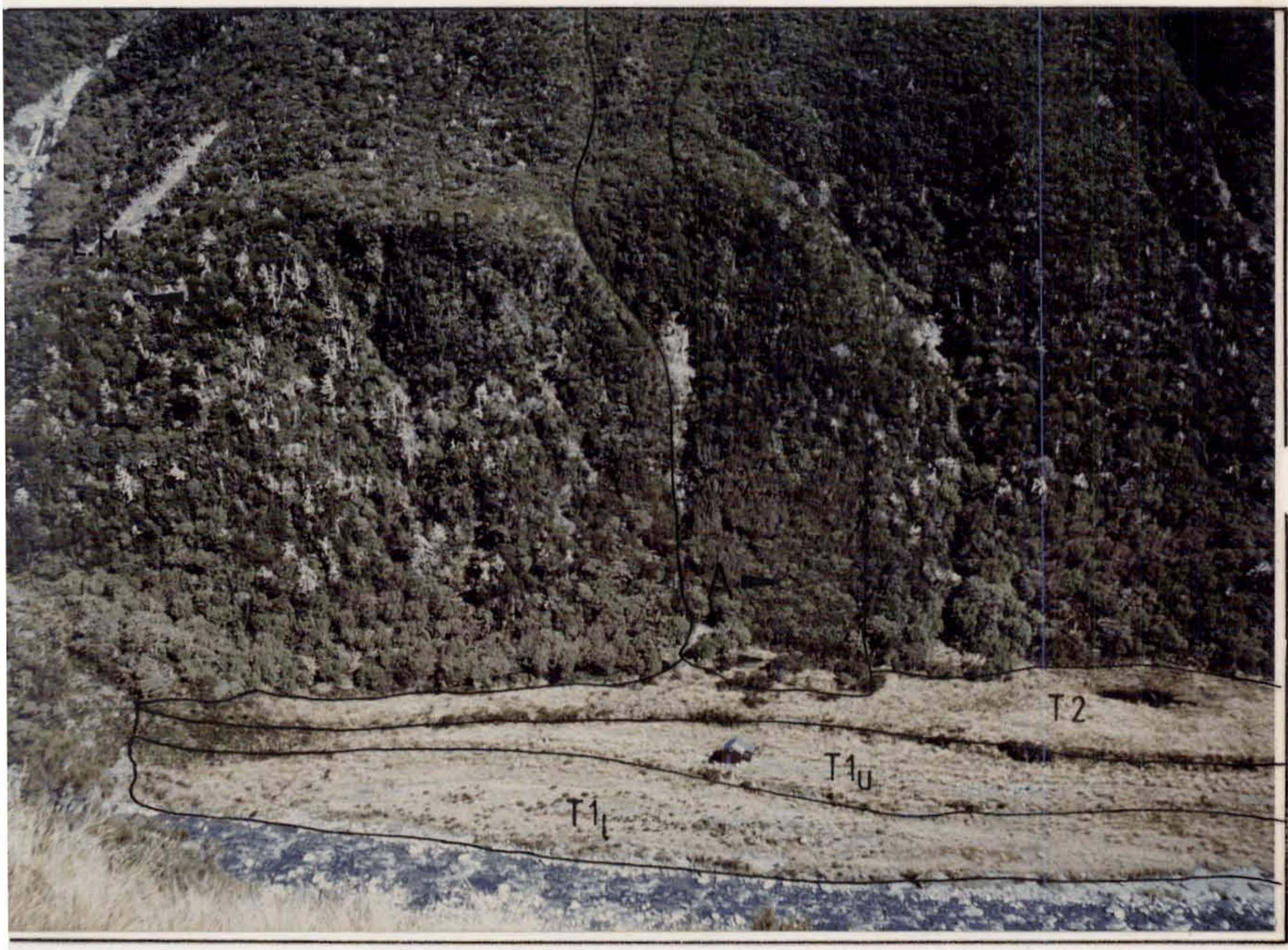
A	F1	I	S1
B	F2	J	S2
C	F3	K	$S3_u$, $S3_l$
D	F4	L	S4
E	F4p	M	$S5_u$, $S5_l$
F	F5	N	S6
G	F6	O	S7
H	F7	P	S10

B.B. - bedrock bench (correlative of Tarkus Knob)

L.M. - lateral moraine


North
O
G
M
Peat Creek
Danger
L
F
4
1
K
6
B
10
H
8
Knob
1
Cropp
River
I
2
J
N
B.E.

Fig. 38: Oblique aerial photo of north bank of Cropp River at Hut Flat showing depositional landforms (see text for key to units T1 to T8.)



D.A. 140 year old debris avalanche

B.B. Bedrock bench (correlative of Tarkus Knob)

L.M. Lateral moraine

Radiocarbon dating site

1 NZ 6347(< 250 yrs BP)

Soil sampling sites

A S2

B S6

Fig. 39: Oblique aerial photo of south bank of Cropp River at Hut Flat.

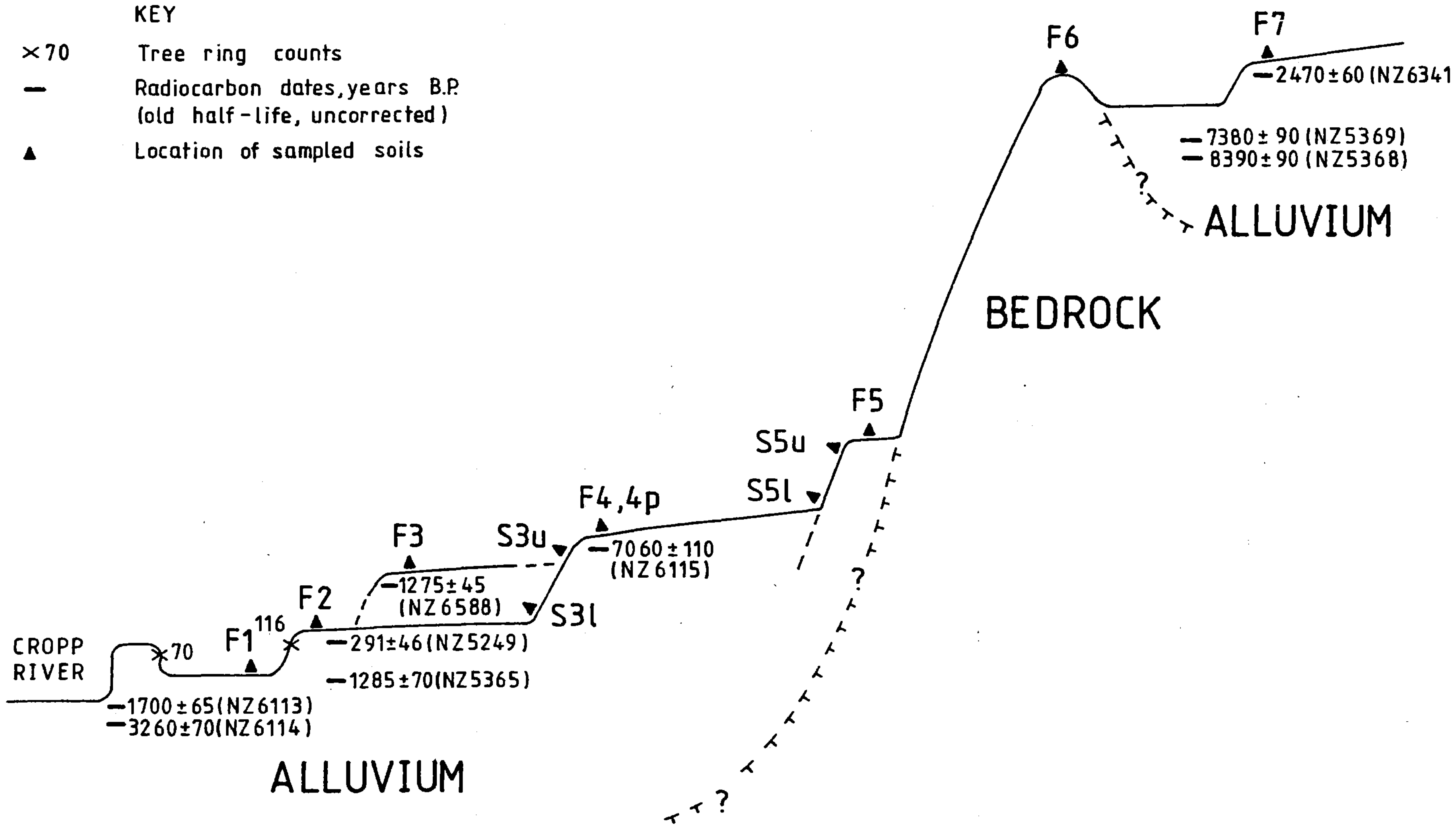


Fig. 40: Stratigraphy of Hut Flat landforms and soils (not drawn to scale)

Location	Species	Growth rings
Riser betwen aggradation and degrational terraces of T1	*Coprosma rugosa*	47
	Coprosma rugosa	51
	Coprosma rugosa	70
Riser between T1 and T2	*Dracophyllum longifolium*	116
Recent debris slide on Tarkus Knob	*Dracophyllum longifolium*	29
	Dracophyllum longifolium	32
	Olearia colensoi	30
Site of S1	*Dracophyllum longifolium*	28
	Dracophyllum longifolium	25
	Olearia lacunosa	29
Debris avalanche behind Cropp Hut, S2	*Dracophyllum longifolium*	121
	Dracophyllum longifolium	134
	Dracophyllum longifolium	125
	Olearia lacunosa	102
	Libocedrus bidwillii	93
	Libocedrus bidwillii	96
	Libocedrus bidwillii	94
	Libocedrus bidwillii	97

TABLE 5: Growth ring counts, Hut Flat (counts made in 1983)

Lab No.	Age (years B.P.)	Location	Material and significance
6318	<250	Hut Flat - base of Tarkus Knob	Wood, in debris avalanche; avalanche younger than 250 years.
6347	<250	Hut Flat - behind Cropp Hut	Wood in debris avalanche; avalanche younger than 250 years (140 ± 5 years on tree ring counts).
6600	<250	Hut Flat	Wood in debris cone; minimal date for debris cone.
5249	A 291 ± 46 B 299 ± 47 C 383 ± 47	Hut Flat, Peat Creek	Wood in alluvium; maximum for surface of T2 and soil F2.
5252	A 363 ± 56 B 373 ± 58 C 435 ± 58	Lower Cropp	Wood in rock avalanche; dates avalanche.
6319	A 1045 ± 60 B 1075 ± 65 C 1025 ± 65	Hut Flat - base of Tarkus Knob	Wood in debris avalanche; dates soil S4.
5250	A 1080 ± 60 B 1110 ± 60 C 1065 ± 75	Hut Flat, Peat Creek	Wood in alluvium; indicates duration of aggradation that formed alluvial cone, T2.
5251	A 1120 ± 60 B 1150 ± 65 C 1100 ± 70	Lower Cropp	Wood in debris cone; dates young episode of aggradation from Knob Creek.
6595	A 1150 ± 60 B 1185 ± 60 C 1130 ± 60	Highest terrace between Reckless Torrent and Snowy Stream	Top of first buried peat layer in alluvium; dates last episode of aggradation at this site.
6587	A 1215 ± 55 B 1250 ± 60 C 1195 ± 60	Cropp River/Reckless Torrent confluence middle terrace	Peat in alluvium; minimum age for formation of terrace.
6588	A 1275 ± 45 B 1310 ± 45 C 1245 ± 55	Cropp River/Danger Gully confluence	Peat in alluvium; minimum age for T3 and soil F3.

Continued next page.

TABLE 6: Radiocarbon dates from Cropp River.

	C 1260 ± 80	Creek	initiation of aggradation that formed T2.
6113	A 1700 ± 65 B 1750 ± 65 C 1685 ± 70	Hut Flat	Peat in alluvium; maximum for initiation of aggradation that formed T2.
5367	A 1750 ± 45 B 1800 ± 45	Hut Flat, top of Tarkus Knob	Basal peat over bedrock; minimum age for alluvial cones on eastern end of Tarkus Knob.
6317	A 2370 ± 70 B 2440 ± 70 C 2510 ± 70	Top Cropp	Basal peat over bedrock; minimum age for deglaciation of head of Cropp River.
6341	A 2470 ± 60 B 2540 ± 70 C 2610 ± 90	Hut Flat, top of Tarkus Knob	Wood in silt; dates soil F7.
6598	A 2930 ± 70 B 3010 ± 70 C 3170 ± 110	Highest terrace between Reckless Torrent and Snowy Stream	Top of lowermost buried peat in alluvium; with NZ6576 provides estimate of age of soil at bottom of pit.
6114	A 3260 ± 70 B 3360 ± 80 C 3550 ± 90	Hut Flat	Peat in alluvium; with NZ6113 indicates duration of degradation by Cropp River.
6585	A 4230 ± 80 B 4350 ± 80 C 4900 ± 160	Noisy Creek	Base of surficial peat; with NZ6348 provides estimate of duration of debris cone deposition.
6115	A 7060 ± 110 B 7270 ± 110	Hut Flat	Base of surficial peat; dates soil F4 and alluvial cone T4.
5369	A 7380 ± 90 B 7600 ± 100	Hut Flat, top of Tarkus Knob	Peat in alluvium; dates stable phase following deglaciation of of Hut Flat.
6348	A 7780 ± 110 B 8000 ± 120	Noisy Creek	Basal peat in colluvium; minimum age for deglaciation of Noisy Creek.
5368	A 8390 ± 90 B 8640 ± 100	Hut Flat, top of Tarkus Knob	Peat in alluvium; minimum for deglaciation of Tarkus Knob.
6576	A 10250 ± 150 B 10550 ± 200	Cropp River, opposite Fan Creek	Wood in till; dates glacial advance.

A Results calculated with respect to old half-life (5568 years).
N.B. These dates are used throughout the text.
B Results calculated with respect to new half-life (5730 years).
C As for B and applying a secular correction (Michael and Ralph, 1972).

contained in Appendix 2. Chemical and physical properties of the sampled soils are described in Chapter 6 while detailed description of the vegetation communities growing on the sampled soils is included in Chapter 7.

5.3.1 Depositional landforms and associated soils

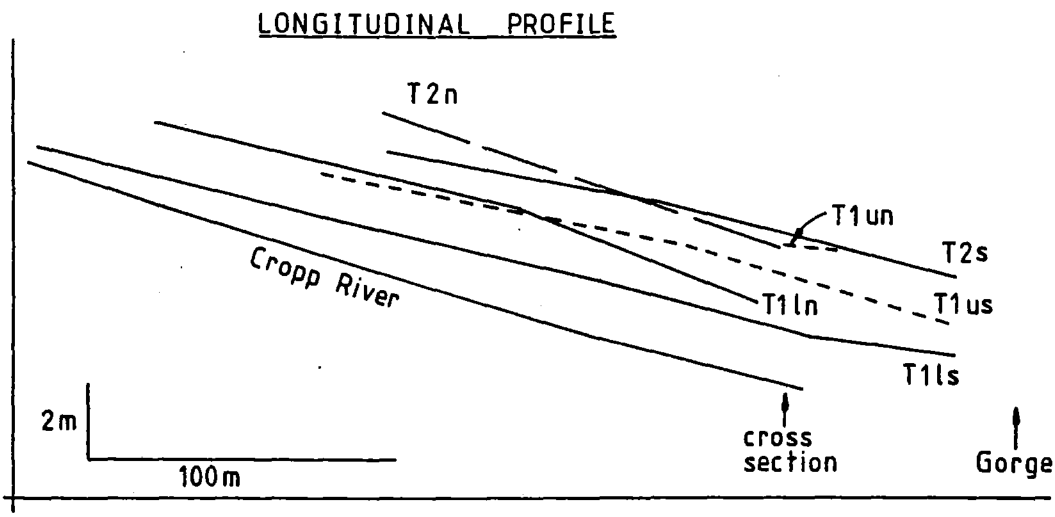
Depositional landforms are characterised by the presence of discrete physiographic units that most nearly qualify as geomorphic surfaces. Many however are time transgressive as a result of fluvial erosion and deposition. The landforms of differing age (T1 to T8), have a complex elevational and spatial zonation.

a) T1 - Very young alluvial terraces.

These occur on both banks of Cropp River and are distinguished by gentle down-valley gradient, incipient soil formation and distinctive seral grassland vegetation. A series of unpaired terraces are present (Fig. 41) that represent both aggradational and degradational phases of terrace formation.

Two terraces are present on each bank (see Figs. 38 and 39). The higher ones are very bouldery and probably are aggradation terraces. The lower terraces commonly have a thin, sandy top stratum over bouldery alluvium, and are degradation terraces. On the south bank of Cropp River the higher terrace is concordant with the surface of a debris cone built by Hut Creek (Fig. 42). This suggests a causal relationship between debris from Hut Creek blocking the course of Cropp River, and aggradation across Hut Flat. Subsequent river incision, and migration across the valley during downcutting, left the series of unpaired terraces.

Terraces mapped as T1 have very weakly developed recent soils. Horizon differentiation is limited to a thin (2 cm), sandy loam textured A_{hj} horizon over unweathered, stony alluvium. (A_{hj}/C horizon sequence). Variation in parent material texture (thickness of sandy top stratum) and site drainage



(Vertical exaggeration =12·5)

KEY
u upper terrace
l lower terrace
s south bank
n north bank

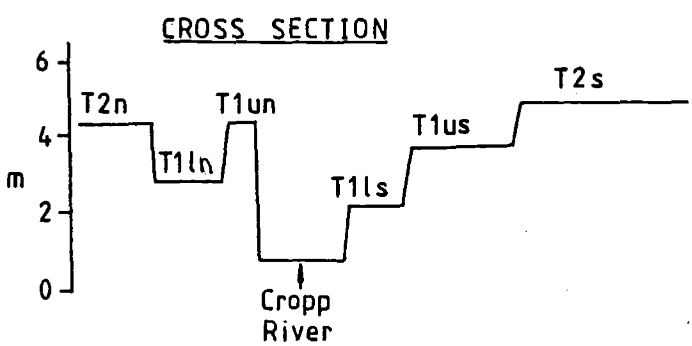


Fig. 41: Surveyed profiles of young alluvial terraces at Hut Flat.

T1 - accordant with youngest, poorly vegetated debris cone from Hut Creek.

T2 - accordant with older tussock and scrub covered debris cone.

Fig. 42: Relationship of T1 and T2 to debris cones formed by Hut Creek.

cause limited soil variability. In wet sites decomposition is retarded and organic soils (with variable amounts of mineral material) are developing. The sampled soil (F1) was located on the lower terrace on the north bank of Cropp River, and had a thin sandy top stratum over bouldery alluvium.

On the north bank of Cropp River, three shrubs on the riser between upper and lower terraces gave ages ranging from 47-70 years (Table 5). This provides a minimal age for the formation of T1. A maximum age of 116 years is likely based on the age of a shrub growing on the riser between T1 and T2. Soil F1 post-dates the formation of the lower terrace and is therefore likely to be less than 70 years old.

b) T2 - Young terraces and alluvial cones

Slightly higher terraces and alluvial cones, with distinctive vegetation and soils, indicate an older phase of aggradation. On the south bank of Cropp River, T2 is the highest alluvial terrace and it is concordant with an older debris cone built by Hut Creek (Fig. 42). T2 is currently accumulating sediment in many places by fluvial deposition and debris avalanching from the hillslopes above.

On the north bank of Cropp River, T2 consists of low angle (2-5°) alluvial cones built by small Cropp tributaries (Peat Creek and Deep Creek). These cones comprise both mineral and organic material. The surface of the alluvial cone built by Deep Creek is dominantly mineral material, while that of Peat Creek is dominantly organic material. In places both cones are affected by recent deposition. The stratigraphy of the Peat Creek cone is exposed where Deep Creek has incised nearly 3 m into it (Appendix 1 - NZ 5249, 5250, 5365). Stratified and interfingered peat and gravel lenses record a complex history of alluvial deposition and peat formation during the period c. 1300-300 years B.P.

Soils on T2 are recent, but have well developed A_h horizons compared to T1. A 10 cm thick, silt loam textured A_h horizon overlies a stony sandy loam textured C horizon containing a few coarse mottles. Variation in parent material texture, age of surface and site drainage causes variability in soils formed on T2, although all soils are recent (A_h/C, A_{hj}/C, O/C horizon sequence). A representative soil (F2, Appendix 2), formed from bouldery alluvium, was sampled from the distal end of the Deep Creek cone.

The age of the surface of T2, and hence of soil F2, is constrained by the following:

(i) a minimum age of 116 years from growth ring counts (Table 5) of a shrub from the riser between T1 and T2 (only the largest shrub was counted as very few shrubs grow on this riser).

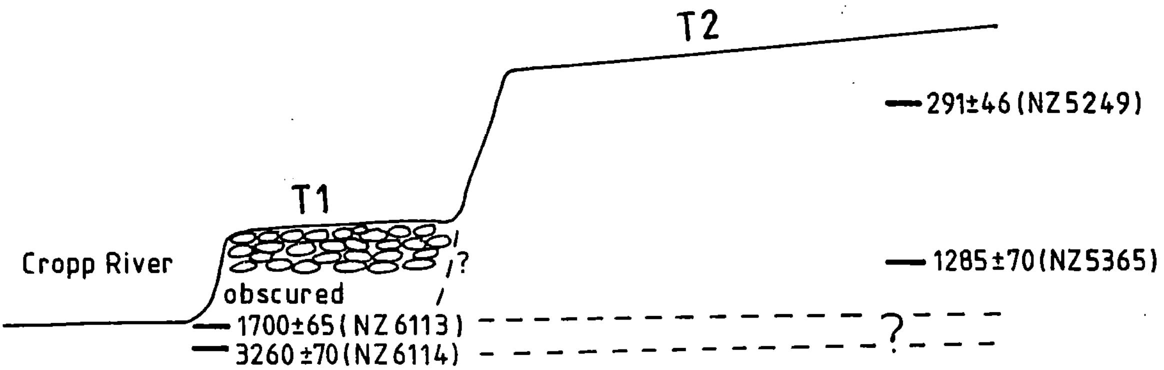
(ii) the date of 291 ± 46 years B.P. (NZ 5249), from 80 cm below the surface of T2 provides a maximum age.

The age of soil F2 is likely to be 200 ± 50 years. This phase of aggradation was initiated at least as early as 1285 ± 70 years B.P. (NZ 5365). The date of 1700 ± 65 years B.P. (NZ 6113) may provide a maximum age for the beginning of aggradation, although the stratigraphic relationship of NZ 6113/6114 to NZ 5365 is not entirely clear (Fig. 43).

c) T3 - Medium age alluvial terrace

On the east bank of Danger Gully is a terrace that is higher in elevation than T2, is more steeply sloping (5-10°), and covered by more mature vegetation and soils. Only a single example of this terrace was identified.

Soils on T3 are far more developed than those on T1 or T2 and have an A_h/E_{agj}/B_w/C horizon sequence, showing characteristics of podzolisation and gleying. A 12 cm thick, silt loam textured A_h horizon overlies 8 cm of grey, silt loam textured E_{agj} horizon with a few mottles. There is a sharp boundary at the base of the E horizon to a thin (1 cm) sandy loam textured B_w horizon, which in turn is underlain by unweathered alluvium. Soils on this terrace are reasonably uniform although there are variations in parent material texture and



Surface of T2 younger than 291 $\pm$ 46 years B.P. Base of T2 underlain by NZ 6113 and 6114 and therefore aggradation of T2 began between 1285 $\pm$ 70 years B.P. and 1700 $\pm$ 65 years B.P.

Fig. 43: Inferred relationship of T1 and T2.

surface age. Younger soils are found along the inner edge of this terrace where an ephemeral stream has more recently deposited sandy alluvium. A representative soil (F3, Appendix 2), was sampled from near the distal end of this terrace. Chemical analysis of F3 (Chapter 6) indicates that it is incipiently podzolised, while morphological features (grey E horizon with few mottles, sharp boundary between grey E horizon and underlying thin B_w horizon) suggest that the E horizon may be gleyed. The sampled soil has 40 to 50 cm of stony sandy alluvium over bouldery alluvium and is finer textured than soils sampled on T1, T2, T4 and T5 (see 6.6.8).

A ^{14}C date of 1275 $\pm$ 45 years B.P. (NZ 6588) was obtained from a peat lens within alluvium in the eroded riser to T3, near the confluence with Cropp River. This site is in the poorly drained distal part of T3 and is considered to record episodes of alluvial deposition in an ephemeral stream draining the inner edge of T3. The date is considered a minimum for the formation of T3 and soils on it.

d) T4 - Old alluvial cone

A large, moderately sloping alluvial cone, considerably higher in elevation than T3, with a distinctive mixed grassland/shrubland vegetation, is prominent on the east bank of Danger Gully. Similar features are present on the west banks of Danger Gully and Whinging Creek (Figs. 37, 38). Average slope is about 12°, but is greater at the cone apex (15°+) and decreases to 8° or less at the distal margins.

The soils on this alluvial cone have far greater morphological development than those on T1 to T3. They are dominantly gley podzols with an $A_h/E_{ag1}/E_{ag2}/B_f/2B_s/2C$ horizon sequence and are characterised by:

(i) a grey, silt loam textured and weakly structured E_{ag1} horizon;

(ii) a more gravelly E_{ag2} horizon with colours of lower value and greyer hue compared to the E_{ag1} horizon;

(iii) a sharp boundary to a thin, discontinuous iron pan between E_{ag} and $2B_s$ horizons. This coincides with a textural break (20% clay in E_{ag2}, 5% in $2B_s$ horizon;

(iv) a thin (2 cm) olive brown, sandy loam textured $2B_s$ horizon.

A representative soil (F4, Appendix 2) was sampled from a relatively well drained site on the distal part of the cone.

Soils with gley podzol features were recorded over most of the cone. In the topographically lower areas of its distal zone, soils with greatly thickened organic horizons have formed. A representative soil from the edge of a peat bog, and located within 5 m of F4, was described ($F4_p$, Appendix 2). The transition from F4 to $F4_p$ occured with only a minor change in topography and appeared to indicate drainage differences. Measurements with a tensiometer (in January 1985, at least 6 days after rainfall) indicated that the surface horizons of F4 (A_h, E_{ag1}, E_{ag2} horizons) were saturated with water, as were the organic horizons of $F4_p$. The contrast between F4 and $F4_p$ may not relate to present day drainage conditions. It is a response to conditions at the time of initiation of soil formation on this cone, when there was a greater topographical difference between F4 and $F4_p$. In the early stages of soil development there were sufficient differences in site drainage to retard organic matter breakdown in $F4_p$ but not in F4. The transition from F4 to $F4_p$ is not marked by any distinctive vegetation changes. Additionally it was observed that there were rarely any soil differences associated with the change from grassland to scrub on this cone. Further variation in soils on T4 is limited to more recent soils found near the eastern edge of T4. Buried soils with a more recent, stony layer over a gley podzol were observed in two places near the cone edge.

A minimum estimate for the age of T4 is provided by a ^{14}C date (7060 $\pm$ 110 years B.P., NZ 6115) from the boundary between the organic and mineral horizons of $F4_p$.

e) T5 - small terrace(?) remnant

Separated from the apex of T4 by a short steep riser is a small, flat-topped bench covered in mixed, low-growing shrubland-grassland. Because of the small size of this feature, and lack of preservation of similar features elsewhere, no unequivocal interpretation is possible. It is however formed of bouldery alluvium and is a remnant of an alluvial terrace or cone.

A soil (F5, Appendix 2) was sampled from the gently sloping summit of T5. Its horizon sequence ($A_h/E_{ag1}/E_{ag2}/B_f/2B_s/2C$) is similar to F4 but it shows an increase in soil depth and the thickness of individual horizons (especially the B_s horizon), colours of greyer hue in the E horizon, and a more continuous and cemented iron pan (B_f).

No direct means of dating this bench is available, but based on profile morphology and its occurrence at higher elevation than T4, it is older than T4 (> 7000 years B.P.). Local glacial history (see 4.2) suggests it is unlikely to be more than 10,000 years old.

f) T6 - Tarkus Knob and remnants of equivalent surfaces

This unit contains the bedrock surfaces on the south-west edge of Tarkus Knob, and small benches at similar elevations on both sides of Danger Gully, and on the hillslopes on the south bank of Cropp River (Figs. 37, 38, 39). These are bedrock surfaces that were overriden by ice during the late Pleistocene (Kumara 3) and early Aranuian advances of glaciers in Cropp basin. A soil (F6, Appendix 2) was sampled from the moderately sloping shoulder of the small bench on the east bank of Danger Gully. It is formed in bedrock but has a horizon sequence ($A_h/E_{ag}/2E_{ag}/2B_s/2C/R$) that is broadly similar to the soils derived from alluvium. Morphological features (in terms of horizon distinctness and thickness) are less well developed than in F4 and F5, and no iron pan is present. This small bedrock bench is substantially eroded around its margins and on topographically higher areas, as are the bedrock surfaces on Tarkus Knob. Thus this unit contains an array of soils from recent to gley podzols.

On the inner edge of the bedrock benches on the north bank of Cropp River is a hollow,that may have been an ice-marginal river channel when Hut Flat was occupied by ice. In Chapter 4 it was suggested that ice filled Hut Flat to about the level of Tarkus Knob at 10,250 $\pm$ 150 years B.P. The bedrock surfaces on Tarkus Knob are therefore unlikely to be older than this date. Dates on peat in the alluvium filling the hollow on the inner edge of Tarkus Knob (see Fig. 37) of 7380 $\pm$ 90 years B.P. (NZ 5369) and 8390 $\pm$ 90 years B.P. (NZ 5368) post-date deglaciation. Thus the bedrock benches are between 8390 and 10,250 years old and F6 is considered to be 9000 $\pm$ 1000 years old.

g) T7 - Alluvial cones on Tarkus Knob

Tarkus Knob has been modified by alluvium filling the hollow described above. The peats from which NZ 5368 and 5369 were obtained record episodes of stability during the development of alluvial cones in the early Aranuian. Few surface features are this old. Younger alluvial cones have been formed by Deep Creek and smaller streams draining the hillslopes above Tarkus Knob. A soil (F7, Appendix 2), on one of these cones showed the maximum morphological development of any soil observed in Cropp basin. It is formed in fine textured alluvium and has a gley podzol horizon sequence ($O_h/A_h/E_{ag1}/E_{ag2}/B_f/B_s/C$) similar to F4 and F5, but with a peaty surface horizon. A thick (30 cm) organic horizon overlies 13 cm of grey, silt loam textured E_{ag} horizon with a sharp boundary to a thin iron pan. Underlying the iron pan is a thin (6 cm), olive brown B_s horizon. Twigs from the E_{ag} horizon gave an age of 2470 $\pm$ 60 years B.P. (NZ 6341). It is considered that the date is a reliable estimate of the age of F7 and that the degree of soil development on a relatively young surface is a result of the fine-textured parent material. Other studies in Westland also show extremely rapid rates of soil development where parent materials are fine textured (Smith and Lee, 1984; Sowden, 1986).

Cone development and peat formation is occurring at present on the eastern end of Tarkus Knob. This area is accumulating peat in the topographically low areas, and mixed mineral/organic material in the better drained areas near the cone apex. A sample from the base of the peat (1.75 m below the surface), near the eastern extremity of Tarkus Knob, gave an age of 1750 ± 45 years B.P. (NZ 5367), indicating that the alluvial cones are less than 1750 years old. Soils on this end of Tarkus Knob are recent (A_h/C) or organic (O/C).

h) T8 - North bank debris cones

Between T4 and Deep Creek, are a series of debris cones at the base of steep hillslopes. They are concave in profile, with slopes ranging from 15-20° at their distal ends, to 30°+ where they merge with the hillslopes. Several small streams cross the cones.

Soils on these debris cones are formed in relatively fine textured, sandy gravelly colluvium and include recent soils (near stream channels and cone apex), weakly gleyed and podzolised soils, and organic soils. Composite soils with weakly developed buried soils, were also common.

It has not been possible to determine the age of these cones with any precision. A radiocarbon date on twigs exposed in the riser to these cones was modern (NZ 6600), showing that they are still accumulating. Stratigraphically they are older than T2 and younger than T4 (i.e. >300 and <7000 years). No soils comparable in development to F7 were observed on these cones, suggesting that all soils on T8 were younger than 2500 years.

5.3.2 Erosional landforms and associated soils

Distinctive erosional landforms corresponding to the age sequence of depositional landforms were not observed on the steep hillslopes surrounding Hut Flat. This is because the majority of older facets of hillslopes have been destroyed by subsequent erosion. However it was possible to select eight soils that could be dated, either directly or stratigraphically. Six of these soils have comparable site conditions and establish changes in soils with time on steep slopes. The other two soils provide information on the influence of slope position on soil development. Additionally four commonly occurring soils were sampled.

5.3.2.1 Soil sequence on steep slopes

a) S1

This was a recent soil formed from debris at the base of a young debris avalanche scar (Fig. 37) where the slope angle was steep (30°). Vegetation comprises open, low (1-2 m), seral mixed scrub. Soils are recent, with horizon differentiation limited to a thin (2 cm) litter layer over colluvium (L/C horizon sequence). Growth ring counts on three shrubs gave ages ranging from 25-29 years (Table 5). A debris slide at the base of Tarkus Knob is of a similar age with shrubs aged from 29 to 32 years. Vegetation on this slide is also low (1-2 m), seral scrub, and soil development is minimal (L/C).

b) S2

This soil was sampled from an older debris avalanche, located directly behind Cropp Hut (Fig. 39). The avalanche is long (c. 150 m) and narrow (20-30 m), with an estimated 1-2 x 10^4 m^3 of debris involved in the failure. The upper part of the avalanche scar is almost vertical, with little colluvium remaining, while the lower part has a prominent footslope and toeslope where much of the debris has come to rest. Tall (4-6 m), seral scrub covers the site.

S2 is a recent soil (H/C horizon sequence) formed from relatively fine textured colluvium (<30% gravel in the C horizon), and occurs in a footslope position with a 29° slope. Horizon differentiation is limited to 5 cm of well decomposed litter overlying colluvium.

Twigs, dated from a toeslope site, gave a modern date (<250 years, NZ 6347). Growth ring counts on various shrubs from this site (Table 5), gave ages ranging from 93 to 134 years, with the greatest ages obtained from *Dracophyllum longifolium* (121-134 years). These ages appear to provide a reliable estimate of the age of the avalanche since *D. longifolium* is an early coloniser, and is prominent on S1. The age of the avalanche is estimated to be 140 ± 5 years.

c) $S3_u$ and $S3_l$

These two soils were sampled from the riser between T2 and T4. This is a short (5-7 m), steep (40°+) rectilinear slope, formed as the edge of an alluvial cone (T4) was eroded by Peat Creek during the formation of another alluvial cone (T2). The soils are formed in colluvium derived from bouldery alluvium. Vegetation ranges from tall (4-5 m) mixed subalpine scrub at the sampling site to short, seral scrub and *Hoheria* towards the apex of T2.

The sampled soils are typical of the most developed soils on this riser. One soil was sampled from the upper backslope ($S3_u$), and the other from the lower backslope ($S3_l$). Both soils are yellow-brown earths (A_h/B_w/2BC/2C horizon sequence). They have 7 to 10 cm of silt loam textured A_h horizon overlying 20 to 25 cm of olive brown, silt loam textured B_w horizon. This is underlain by a stony sandy loam textured BC horizon that is transitional to the parent material. The main difference between the two is in B_w horizon colour (2.5Y 5/6 in upper, and 2.5Y 4/3 in lower backslope) and the presence of mottles in $S3_l$. Less developed soils (A_h/C, L/C horizon sequence) occur on this riser toward the apex of T2, and where ephemeral stream channels cross the riser.

The age of the riser can be inferred from the history of T2 presented in 5.3.1b. It was suggested that aggradation began between 1285 ± 70 years B.P. (NZ 5365) and 1700 ± 65 years B.P. (NZ 6113), and continued to between 116 and 291 ± 46 years B.P. (NZ 5249). This provides bracketing ages for $S3_u$ and $S3_l$ (riser age is less than 1700 years B.P. and greater than 116 years). These soils have a lesser degree of profile development than S4, which is directly dated at about

1000 years B.P., and they have been assigned an age of 500-1000 years. However, the parent material of S4 is finer textured than that of $S3_u$ and $S3_l$, and this may have affected the rate of soil development.

d) S4

This soil was sampled from a steep (44°), colluvium covered hillslope near the base of Tarkus Knob, (Fig. 37) and is directly dated by a buried soil in the footslope below the soil sampling site. Vegetation on this site is tall (4-6 m), mixed scrub with emergent *Libocedrus bidwillii*.

S4 is a podzolised and gleyed yellow-brown earth ($A_h/E_{agj}/B_s$/2BC/2C horizon sequence). It has a thin (3 cm), silt loam textured A_h horizon, over a thin, variable (5 to 12 cm), greyish E horizon with a few mottles. There is a clear boundary to a thick (65 cm) olive brown, silt loam textured B_s horizon which is underlain by 40 cm of material transitional to the C horizon. Incipient podzolisation is evident, both morphologically (thin, variable, greyish E_{agj} horizon) and chemically (Chapter 6), while gleying is indicated by the dominant greyish matrix colour in the E_{agj} horizon with a few mottles.

A pit in the footslope below the sampled soil revealed three buried soils. ^{14}C dates (NZ 6318, 6319) were obtained from wood in the upper two soils. Both soils relate to periods between episodic debris avalanching. The modern date relates to a small failure at the base of the slope, while the date of 1045 $\pm$ 60 years B.P. provides a maximum age for the deposits in which S4 has formed.

e) $S5_u$ and $S5_l$

Two soils were sampled from the riser between T4 and T5. This short, steep (40°+) rectilinear slope is very similar to that on which $S3_u$ and $S3_l$ were sampled, except for its greater age. Its age is 7000 to 10,000 years since it pre-dates the formation of T4, and post-dates glacial occupation of Hut Flat.

There is no evidence of erosion or deposition at the sampling site and the degree of soil development and catenary relationships suggest this short slope has been stable since formation.

Two soils, from the upper and lower backslope, were described and analysed and another from mid-backslope was described. These three soils form a catena with a podzolised yellow-brown earth (A_h/B_s/BC/C/2C/3C) in the upper backslope ($S5_u$), a gley podzol ($A_h/E_{ag}/B_s$/BC/C) in the mid-backslope (similar to S6), and a gley soil (H/B_g/C) in the lower backslope ($S5_l$). $S5_u$ consists of 10 cm of silt loam textured A_h horizon, over a thick (50 cm) yellowish brown, silt loam textured B_s horizon. This is underlain by 35 cm of material transitional to the parent material. $S5_l$, by contrast, has a thick (16 cm) organic surface horizon, overlying a thick (107 cm) B_g horizon with dominant grey colours.

The vegetation comprises tall (4-5 m), mixed scrub with emergent *L. bidwillii* and *Dracophyllum* sp., similar to that on S6. No gross vegetation differences occur with slope position.

f) S6

This soil was sampled from under a mature, species rich, subalpine scrub-conifer forest on a stable, steeply sloping (35-45°) ridge about 75 m above Hut Flat (Fig. 37). It was chosen because it showed the most advanced soil morphological development on steep slopes. Its age is unknown but it is on a surface which was overriden by ice in the early Aranuian. It is unlikely to be >10,000 years B.P. and could be significantly younger.

The surface has a stepped microtopography and the soil was sampled from a 27° step. It has a gley podzol horizon sequence ($H/A_h/E_{ag}/B_s$/2BC/2C). A thick (20 cm) surface organic layer overlies 5 to 15 cm of bluish grey, silt loam textured E_{ag} horizon with massive structure. There is an abrupt boundary to 35 cm of olive brown, silt loam textured B_s horizon which gradually gives way to material transitional to parent material. S6 is similar to the older soils on gentle slopes except for an organic surface horizon, lack of an iron pan, a more gradual boundary between the E_{ag} and B_s horizons, and a thicker B_s horizon.

In January 1985 (at least 6 days after rain) the soil water status of this soil was investigated. Tensiometer measurements indicated saturation to a depth of at least 1 m, and no differences between gleyed horizons (E_{ag}) and horizons showing no morphological evidence of gleying (B_s). These measurements, and the known frequency of rainfall, suggest that, even on steep slopes, the soils remain saturated throughout the profile for most of the year. The consequences of soil hydrology to soil genesis are discussed in Chapter 8.

<u>5.3.2.2 Other commonly occurring soils</u>

Three shallow, recent soils and one shallow, old soil were sampled. These profiles are representative of some of the more common soils in Cropp basin.

The 3 recent soils (A_h/C horizon sequence) are characteristic of the extensive areas of steeply sloping (40-50°), highly dissected, tussock grasslands, at higher elevations, underlain by pelitic schist. One soil (S7) was sampled from Hut Flat (Fig. 37) and two (S8, S9) from Danger Gully (Fig. 24). The latter two were sampled as part of an investigation of the relationship between soil development, soil nutrient status and plant element concentrations (see Chapter 7) and are typical of soils in the pelitic schist region. These recent soils are among the most common, in areal extent, in Cropp basin. S7 and S8 have bedrock within 20 cm of the surface while S9 has 40 cm of colluvium over bedrock. By extrapolation from the dated soil sequences all three soils are estimated to be 100-200 years in age.

Shallow gley soils occur wherever soils are forming directly on bedrock resistant to weathering. S10 is typical of the older soils occurring in such sites. It is formed on an 18° slope underlain by resistant quartzose schist, with low (1-2 m), open shrubland vegetation. It has 10 cm of dark grey, silt loam textured A_h horizon overlying 10-20 cm of bluish grey, silt loam textured E_{ag} horizon with a sharp boundary to bedrock. No illuvial horizons occur, although there is an incipient iron pan in places at the E_{ag}/R boundary. No direct date is available for this soil, but it is on a surface which was probably formed during the early Aranuian glacial advance in Cropp basin. It is therefore younger than 10,000 years, and possibly considerably younger.

5.3.2.3 Geomorphology and soil patterns of steep slopes

Geomorphology is discussed with reference to slope aspect because there are significant differences between north- and south-facing hillslopes at Hut Flat (Figs. 38, 39). This difference primarily results from contrasts in bedrock lithology, and interaction of structure with topography, rather than microclimate. Microclimate does however influence the vegetation pattern (e.g. grassland extends to lower altitudes on south-facing slopes; *Libocedrus bidwillii* is far more prominent on north-facing slopes).

South-facing hillslopes (on the north bank of Cropp River) are underlain by pelitic schist. A broad amphitheatre shape has developed (Fig. 38) with a dense ephemeral drainage network. As well as the prominent fluvial dissection, small shallow mass movements are common, but revegetate rapidly to maintain a dense vegetation cover. Slopes are generally short and steep (30°+), and steepen progressively from valley bottom to ridge. Significant depositional landforms are present at the base of the south-facing hillslopes.

Slopes at the head of the amphitheatre, under tussock grassland, are steep (40-50°) and underlain by shallow recent soils (A_h/C e.g. S7). In the middle part of these hillslopes, a greater depth of colluvium is present (1-2 m +), and it is commonly stratified. Recent soils occur under both tussock grassland and scrub in this zone. Scrub is restricted to steeply sloping ridges, and tussock to hollows where snow lies longest. The lower, more gently sloping (25-35°) part of these hillslopes is covered in a mosaic of scrub and tussock, and is crossed by several small streams. In this area, yellow-brown earths are found in the more stable parts of the landscape (e.g. interfluves) and recent soils on sideslopes adjacent to stream channels. Small remnant slope facets, with gley podzol soils, also occur.

On the SW flank of Tarkus Knob is a steep (c. 40° average) slope formed on resistant bedrock. This long (c. 100 m) slope probably had a glacial origin but has been affected by debris avalanches-slides. The soil pattern is a mosaic of recent soils, yellow-brown earths, podzolised and gleyed yellow-brown earths (e.g. S4), gley podzols formed in colluvium, and shallow soils over bedrock (e.g. S10). This slope is similar to north-facing slopes (in terms of geomorphology and soils) because the underlying geology is psammitic rather than pelitic schist.

North-facing hillslopes (Fig. 39) are long and steep (30-70°). Dissection of these hillslopes is not so prominent although much is hidden by the dense vegetation (see Fig. 26). Numerous small ephemeral streams drain these slopes. Debris avalanches are common, often associated with the stream channels. Scars range in size from large features, such as that on which S2 has formed, to small, narrow features that are barely visible amongst the scrub. Windthrow of individual trees or small groups of trees is also common.

The soil pattern on the north-facing hillslopes is a mosaic of recent soils (S1, S2), yellow-brown earths, podzolised and gleyed yellow-brown earths, gley podzols (S6) and gley soils. More developed soils are found in stable sites on noses and broad ridges, particularly in the higher parts of the landscape. Recent and yellow-brown soils predominate in hollows, on steeper slopes, and in lower slope positions, and cover the majority of the land surface.

Commonly the north-facing slopes become precipitous (50-70°) at the base, as a result of undercutting by streams (Hut Creek and Cropp River). The base of these hillslopes is marked by a prominent strip of *Hoheria* forest (Fig. 39) where a thin veneer of slope wash is accumulating. Only minor depositional features are present compared to those at the base of the south-facing slopes.

5.3.3 Soil development sequences

The array of erosional and depositional landforms at Hut Flat has been used to examine soil development sequences in two contrasting situations:

(a) gentle to moderate slopes (<15°) with soils derived from alluvium and formed under tussock grassland or mixed grassland/shrubland (F1 to F7), and

(b) steep to very steep slopes (30-50°) with soils derived from colluvium and formed under subalpine scrub and scrub-forest (S1 to S6).

In Fig. 44 the sampled soils are arranged according to age, parent material and topography. Fig. 45 summarises the changes in soil profile horizonation with age for both soil sequences. These soil development sequences, and the chronology for soil development have been used to interpret the age of soils elsewhere in Cropp basin.

Soils F1-F5 are members of a post-incisive (Vreeken, 1975a), non-strict chronosequence (Stevens and Walker, 1970) formed from coarse textured schist alluvium. There are minor differences in parent material texture and site conditions (especially topography), but the changes in soil morphology arise primarily from increasing soil age from F1 to F5. Soil development is slower in bedrock (F6 is morphologically less developed than F4 or F5) and more rapid in fine textured alluvium, as illustrated by F7.

Six of the soils on steep slopes (S1, S2, $S3_u$, S4, $S5_u$, S6) also form a post-incisive, non-strict chronosequence, but with

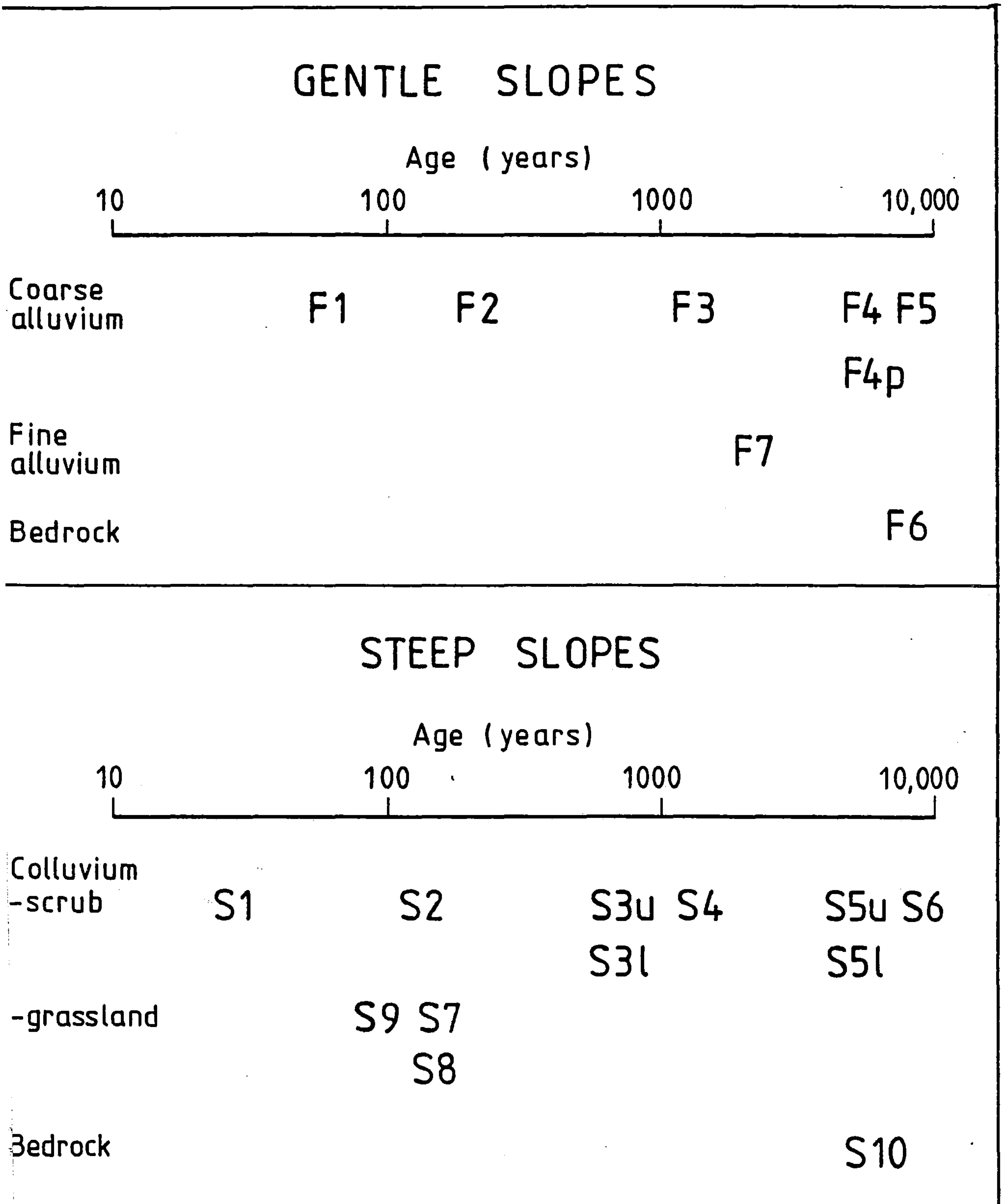


Fig. 44: Sampled soils arranged according to age, parent material and topography.

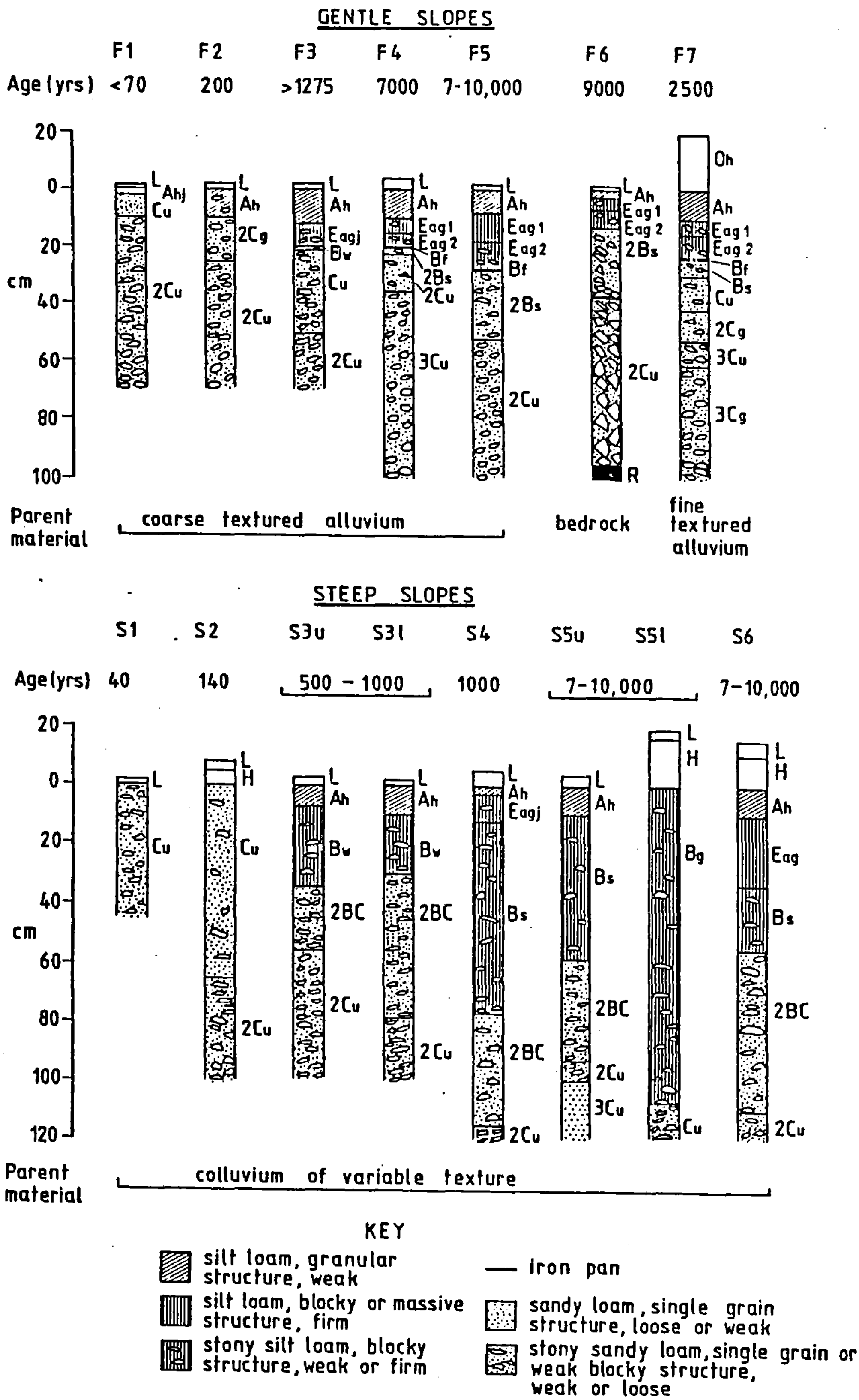


Fig. 45: Changes in soil profile horizonation with age for the soil sequences on gentle and steep slopes.

more variability in site conditions and parent materials:

(a) Parent material texture - includes soils derived from relatively fine textured colluvium and coarse textured alluvium. Gravel content of the C horizons ranges from 32 to 74%;

(b) Topography - slopes range from 27-50° while slope position (which affects the soil water regime) ranges from upper backslopes to footslopes. Aspect also varies - includes soils formed on both north- and south-facing slopes.

In this sequence $S5_u$ is somewhat anomalous as it lacks an E_{ag} horizon, although an E_{ag} horizon was present downslope of $S5_u$ (mid-backslope). Comparison of $S3_l$ and $S5_l$ with $S3_u$ and $S5_u$ indicates the effect of slope position on soil development.

The sequences provide a model of soil development for coarse textured alluvial or colluvial parent materials, in sites where the surface of the soil remains sufficiently aerated to preclude the formation of thick (>30 cm) surface organic horizons. In general terms, both sequences show the following progressive changes in soil profile form:

(a) Recent soils - on surfaces less than 200 years in age (F1, F2, S1, S2);

(b) Yellow-brown earths - on surfaces 500-1000 years in age ($S3_u$, $S3_l$). This phase is not represented on gently sloping sites

(c) Podzolised and gleyed yellow-brown earths - on surfaces 1000-1500 years in age (F3, S4);

(d) Gley podzols - on surfaces 7000 years or more in age (F4, F5, S6).

With increasing age, there are increases in the thickness of individual horizons and in total soil depth, increasing expression of podzol and gley features, and a major change in the texture of surface horizons. Fine earth texture profile form changes from gradational negative to duplex negative in soils from alluvium, or from uniform to gradational negative in soils from colluvium. Redox profile form changes from raw unoxidised in the recent soils, to oxidised in the yellow-brown earths, to middle reduced in the gley podzols. Structure profile form is subpedal in the recent soils and becomes isopedal or bipedal with increasing soil development. Eventually structure in the E horizon becomes massive and an apopedal structure profile form develops.

There are several significant differences between the two soil sequences. Compared to soils on gentle slopes, those on steep slopes had more organic accumulation (as evidenced by the depth of litter layers and organic horizons), greater soil depth and B horizon thickness, and no iron pan between the E_{ag} and B_s horizons. These differences are caused by variation in the state factors biota, parent material and topography. Soils on gentle slopes have formed predominantly under grassland, whereas those on steep slopes have formed under scrub. Soils on gentle slopes have formed from alluvial parent materials, whereas those on steep slopes have formed from colluvial parent materials. The different topographical settings of the two soil sequences affect soil water regimes and hence soil development.

Catenas were present on stable, gently sloping surfaces ($F4/F4_p$) or on steep slopes of short length ($S5_u/S5_l$). On the alluvial cone risers catenary sequences were not present at 500-1000 years ($S3_u/S3_l$), but were well developed by about 7000 years ($S5_u/S5_l$) indicating that at least 1000 years are needed for catenas to develop on steeply sloping topography. Many soils are younger than this, and temporal factors therefore influence soil distribution more strongly than spatial factors. In general few catenary sequences occur.

5.4 LOWER MID-CROPP REGION

This includes the area of Cropp basin between Tarkus Knob and Reckless Torrent/Cream Creek (Figs. 46, 47) where depositional landforms are most prominent. On the south bank of Cropp River large debris cones are present, while on the north bank there is an array of surfaces broadly similar to

A.C. Alluvial cone

Mo Moraine

T_u Upper terrace

T_l Lower terrace

Radiocarbon dating sites

1 NZ 6595 (1150 ± 60 yrs BP), NZ 6598 (2930 ± 70 yrs BP)

2 NZ 6587 (1215 ± 55 yrs BP)

3 NZ 6576 (10 250 ± 150 yrs BP)

Fig. 46: Oblique aerial photo of the north bank of Cropp River between Snowy Stream and Reckless Torrent.

Fig. 47: Oblique aerial photo of debris cones on the south bank of Cropp River between Hut Creek and Cream Creek. Note vegetation pattern.

that at Hut Flat. The north bank was studied in some detail because of its similarities with Hut Flat and the abundance of material for dating.

5.4.1 North bank

An array of terraces and alluvial cones are present between Snowy Stream and Reckless Torrent (Fig. 46). A very young alluvial terrace is present adjacent to both Snowy Stream and Reckless Torrent. It has recent soils (A_{hj}/C horizonation) and is flooded periodically. Vegetation cover is mainly seral grassland dominated by *Chionochloa pallens* and *C. rubra*. The age of shrubs of *Hoheria glabrata* present on this surface were estimated, from diameter, to be 100-200 years.

A slightly higher alluvial terrace is present with more developed soils (A_h/C horizonation) and grassland dominated by *C. rubra*. In a pit on the edge of this terrace, at the confluence of Cropp River and Reckless Torrent, four buried peats, separated by stony or gravelly alluvium, overlie bouldery alluvium. The lowermost peat gave an age of 1215 $\pm$ 55 years B.P. (NZ 6587). This date is very similar to that obtained from the base of T2 at Hut Flat (1285 $\pm$ 70 years B.P., NZ 5365) suggesting possible synchroneity of aggradation at the two sites. The surface of this terrace is now more than 10 m above Cropp River but parts of it are subject to episodic alluvial deposition by small streams (as recorded in the soil pit).

The highest surface between Snowy Stream and Reckless Torrent (30-40 m above Cropp River) is formed on two deposits of greatly different age. The eastern end has an extremely uneven surface topography with many large boulders (up to 10 m diameter) and is a moraine surface. The underlying till was dated at 10,250 $\pm$ 150 years B.P. (NZ 6576). Vegetation cover on this component of the surface is mature, low growing subalpine scrub-conifer forest (see Chapter 7). Soils are commonly strongly developed gley podzols except where the surface has been affected by localised fluvial erosion. In such sites recent and yellow-brown earth soils occur.

The western end of this high surface is distinguished by the absence of large boulders, and by a *Chionochloa rubra* dominated tussock grassland vegetation. This component of the surface comprises small, low angle alluvial cones that have buried the moraine surface.

On these cones recent soils (A_h/C horizon sequence) occur adjacent to stream channels, and composite recent soils further from active channels. At the extreme south-west end is a peat bog consisting of stratified peats and alluvium. This site is beyond the influence of currently active streams. A pit dug in this bog showed three buried peat layers, separated by sandy or gravelly alluvium, overlying a gley podzol. Large boulders, possibly till, formed the bottom of the pit (1.7 m). The top of the lowermost organic horizon (buried topsoil of gley podzol) gave a date of 2930 $\pm$ 70 years B.P. (NZ 6598), while the top of the first buried peat was dated at 1150 $\pm$ 60 years B.P. (NZ 6595). Three major periods of alluvial deposition have occurred at this site in the last 3000 years.

5.4.2 South bank

On the south bank of Cropp River between Hut Creek and Cream Creek are a series of moderately sloping debris cones formed by streams draining Steadman Brow (Fig. 47). These have formed by rockfall, debris and snow avalanches, debris flows and fluvial deposition. One of the cones is partly underlain by glacial and fluvioglacial sediment (see 4.2.2). The debris cones have concave profiles with slope angles increasing from 15-20° in the distal zones, to 30-35° at the cone apices. Bedrock underlies the debris cones, generally within 50 m of the surface.

Four streams, deeply incised in narrow channels, cross these cones. Depth of incision is greatest in the distal zone of each cone (up to 40-50 m) and decreases towards the apex. All four streams join Cropp River discordantly, suggesting downcutting by these tributaries is not as rapid as that of Cropp River.

Distinctive vegetation patterns, caused by stream avulsion on the debris cones, are evident in Fig. 47 and are well correlated with the broad soil pattern. The pattern comprises:

(a) low seral *Dracophyllum-Olearia* scrub on the youngest surfaces adjacent to streams, with recent (L/C or A_{hj}/C) soils;
(b) tall seral *Dracophyllum-Olearia* scrub, with recent soils and yellow-brown earths;
(c) tall, mixed *Dracophyllum-Olearia* scrub with prominent *D. traversii*, and occassional *Libocedrus bidwillii* and *Dacrydium biforme*. Yellow-brown earths, podzolised and gleyed yellow-brown earths are the dominant soils;
(d) mixed scrub-forest with prominent *L. bidwilli* and *D. biforme*. Soils are similar to (c);
(e) scrub-forest, with canopy dominated by *D. biforme* and *L. bidwillii*, occurs with gley podzol soils (comparable to S6).

The Cream Creek site (Fig. 47) described by Norton (1983a, b) falls within vegetation community (d). Soils on this debris cone were mainly yellow-brown earths but included recent soils and incipiently podzolised and gleyed yellow-brown earths (similar to S4). By comparison with the dated soils at Hut Flat, the maximum stage of soil development observed suggests that the Cream Creek cone is 1000-1500 years in age. The oldest tree aged by Norton (1983a) from this site was 600 years old. The surface of the debris cone is now about 40 m above Cream Creek.

The vegetation pattern on the south bank debris cones indicates a dominance of young surfaces and soils. Vegetation communities (a) to (d) are predominant with minor areas of scrub-forest (e). Based on vegetation characteristics most surfaces on these debris cones are younger than the Cream Creek site (<1000 years). The presence of till within one cone indicates that they probably began forming c. 10,000 years ago, and provides a likely maximum surface age.

5.5 NOISY CREEK

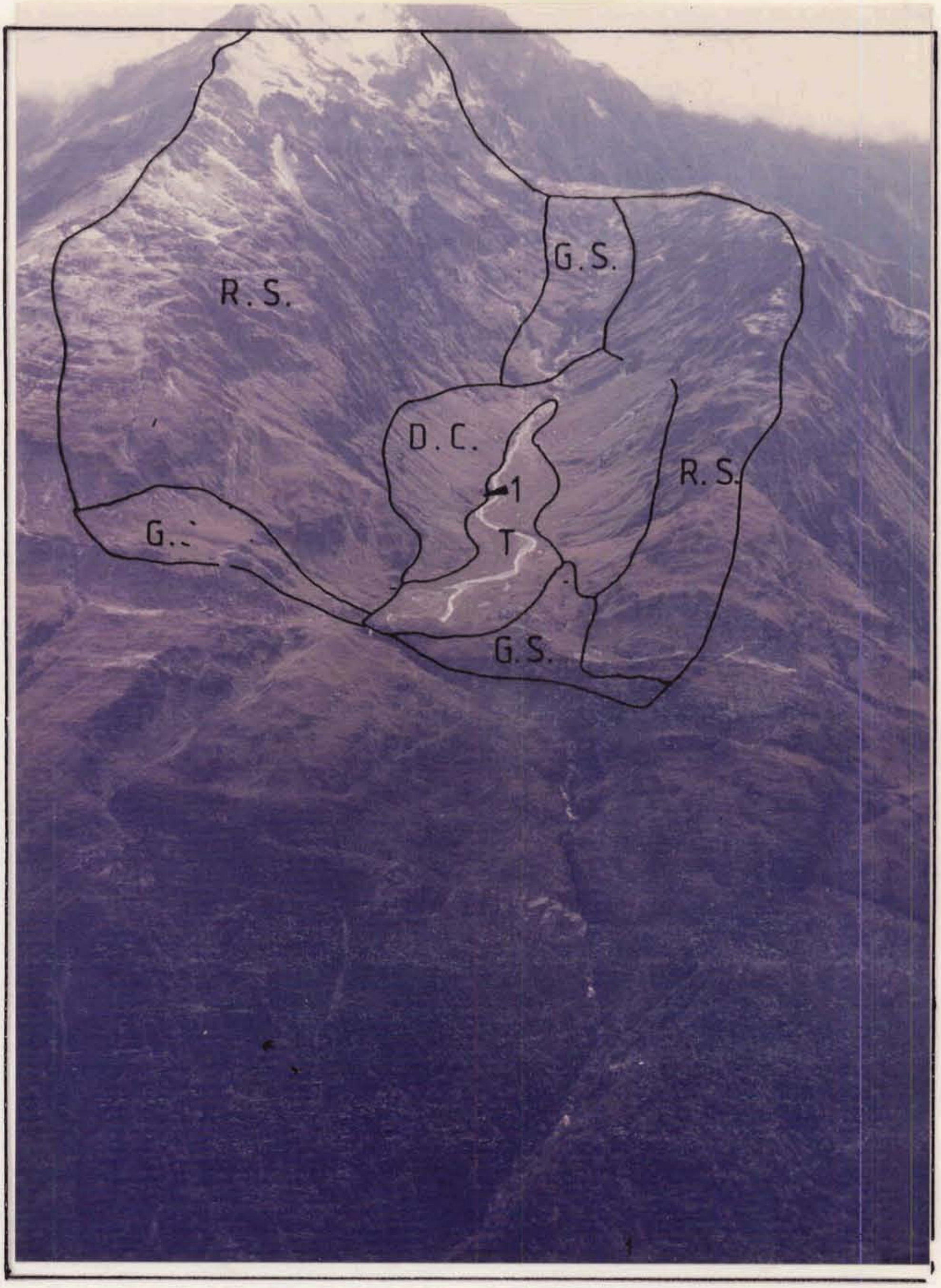
Noisy Creek drains a cirque basin in a hanging valley formed behind an outcrop of ultramafic rocks (Pounamu Formation). The resistance of the Pounamu ultramafics has slowed vertical corrasion by glaciers and rivers and controlled the development of this valley. (see Fig. 48 and frontispiece). Behind the Pounamu ultramafics a cirque was excavated by glacial erosion, and this was occupied by a lake following deglaciation. Below the Pounamu ultramafics, Noisy Creek falls steeply to join the Cropp River (Fig. 35, 48). Only the hanging valley was investigated, as an example of an area where base level had been relatively stable. Noisy Creek exits the hanging valley through a narrow bedrock gorge that is incised 20-30 m into the Pounamu ultramafics.

The effect of glaciers on Noisy Creek is clearly evident in the U-shaped valley, valley-in-valley form and ice scoured and striated bedrock surfaces. Hawkes (1981) inferred from the valley-in-valley form that two glacial events were recorded, which he correlated with the Kumara-3 and Waiho Loop advances. A radiocarbon date of 7780 $\pm$ 110 years B.P. (NZ 6348) was obtained from peat near the base of a debris cone in Noisy Creek, implying that the area was substantially deglaciated by that time.

The following landforms occur in Noisy Creek: young alluval terraces in the valley floor; alluvial and debris cones discordant with the present river level, with truncated distal margins; steep rock hillslopes on basin perimeter; gently to moderately sloping, bedrock surfaces at the head of Noisy Creek, and its exit through the Pounamu ultramafics.

a) Alluvial terraces in valley floor

A series of terraces are present that range from 0.5 to <5



T Alluvial terraces

D.C. Debris and alluvial cones

R.S. Steep rock slopes

G.S. Gently sloping bedrock surfaces

Radiocarbon dating site

1 NZ 6565(4230 ± 80 yrs B.P.), NZ 6348(7780 ± 110 yrs B.P.)

Fig. 48: Oblique aerial photo of Nobby Creek cirque basin.

m above the present channel of Noisy Creek (Figs. 49 and 50). These are almost flat (<2°) at the eastern end of the valley and gently sloping (5 to 7°) towards the valley head. Much of the valley floor is affected by deposition during large floods (Fig. 50). On the lower terraces, soils are recent, (C, A_{hj}/C, A_h/C profile forms), or organic. Higher and older terraces occur towards the eastern end of the valley on the southern bank of Noisy Creek. The higher terraces have a dominance of organic soils, but in freely draining sites weakly gleyed and podzolised soils have formed ($A_h/E_{agj}/B_w/C$ horizonation).

In the same area of the valley well sorted, laminated silts and sands with graded bedding underlie the terrace alluvium in places (Fig. 51). These are lacustrine sediments deposited in a lake that filled Noisy Creek behind the Pounamu ultramafics. The lacustrine sediments were found in 3 sites, under both low and high terraces, near the eastern end of Noisy Creek. Elsewhere they have been eroded by Noisy Creek as its channel incised. No material was obtained for directly dating either the terraces or the lake. Based on soil profile morphology, the terraces are mainly <1000 years old, with a few older remnants. All the terraces post-date the lake.

b) Alluvial/debris cones

At the base of the steep hillslopes that form the perimeter of Noisy Creek are a series of moderately sloping (10-20°) alluvial cones and steeply sloping (20-40°) debris cones. The cone surfaces are discordant with the present valley floor (Figs. 49, 50) and appear concordant with the crest of the Pounamu ultramafics (gradients surveyed using an abney level). Horizontal benches are prominent in the risers formed by the truncated distal margins of the cones, and also in bedrock outcrops.

Two dates were obtained from one debris cone (Fig. 48). Peat, from 4.5 m below the surface and overlying angular, stony colluvium, was dated at 7780 ± 110 years B.P. (NZ 6348). The

Fig. 49: Noisy Creek headwaters - note flat terraces; truncated alluvial and debris cones with horizontal benches (lake shoreline); bedrock surfaces at head of valley with deep ravines.

Fig. 50: Noisy Creek, near exit from cirque basin - note recent deposition on low terraces; truncated debris cones; horizontal benches in debris cones and bedrock.

Fig. 51: Noisy Creek, relationship of terrace alluvium to lacustrine silts. Peat overlies bouldery alluvium, which in turn overlies lacustrine sediments which extend to at least 4 m depth.

base of the surface peat (at 1 m depth) was dated at 4230 ± 80 years B.P. (NZ 6585). These dates indicate the gross form of this debris cone developed between c. 4000 and c. 8000 years ago.

The accordance of cone gradients with the crest of the Pounamu ultramafics suggests the debris and alluvial cone surfaces were graded to a lake level. Modern analogues exist at Ivory Glacier where alluvial cones are graded to lake level and have very steep sublacustrine margins (Robinson, 1981). The Ivory Glacier basin is less than half the area of Noisy Creek but a hollow in excess of 50 m deep has been excavated by glacial erosion. The depth of infilling that has occurred in Noisy Creek is unknown. The lake existed from before 7780 years B.P. to at least 4230 years B.P. The presence of horizontal benches at heights below the crest of the Pounamu ultramafics suggests the lake persisted following significant base level lowering at the outlet after 4230 years B.P.

An array of recent soils, yellow-brown earths and podzolised and gleyed yellow-brown earths occur on the alluvial and debris cones, depending on surface age. Organic soils occur in the wettest areas. Recent soils are associated with the numerous stream channels and upper parts of slopes. Older soils occur on topographically higher areas (noses) and distal parts of debris/alluvial cones. Although the gross form of these cones may be relatively old, the surfaces of most are still receiving sediment and recent soils are predominant.

A single debris cone was selected for study and a series of soil pits were dug along the slope profile. (Fig. 52). This cone is formed at the base of steep (50-90°) rock hillslopes, and grades to a rock bench downslope. This rock bench, which is traceable laterally, may relate to the former lake level. A gley podzol soil (NC1) occurs on the rock bench under sparse, short *Chionochloa pallens* and *C. crassiuscula*. This soil is traceable, laterally towards a stream and upslope, to where it is buried by more recent colluvium (NC2) and where

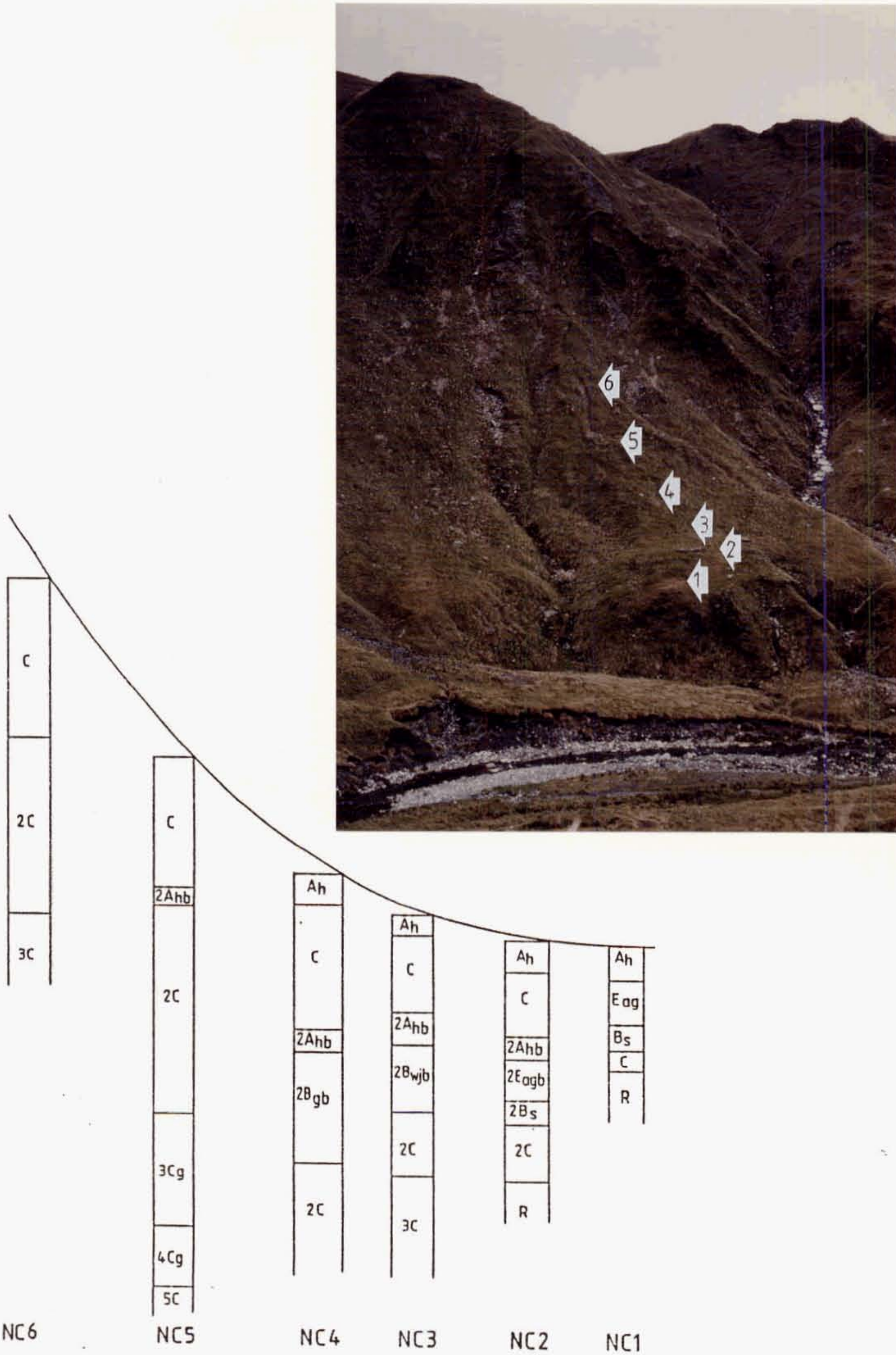


Fig. 52: Study site, debris cone on north bank of Noisy Creek.

vegetation is dense, tall *C. pallens*. Further upslope, all soils are recent with indistinct sedimentary layering (NC3-NC6), indicating frequent, episodic slope failures (debris and snow avalanching, fluvial erosion/deposition) from the steep hillslopes above.

c) Steep, rock hillslopes

Steep (40-90°), sparsely vegetated hillslopes, underlain by bedrock, surround the perimeter of Noisy Creek. These areas have little soil cover and are eroding by fluvial dissection, rockfall and snow avalanching. They are the source areas for the depositional landforms. Numerous large (up to 7 m diameter) boulders on the terrace and cone surfaces have fallen from these steep hillslopes.

d) Gently sloping bedrock surfaces

This includes areas near the basin outlet that were glaciated in late Otiran-early Aranuian advances, and the head of the valley where glacial surfaces are younger.

Immediately upstream of the Pounamu ultramafics, on the north side of Noisy Creek, is an area of rolling topography (0-20°) underlain by psammitic schist. On some convexo-concave slopes catenary sequences of soils have formed. Gley podzols occur on summits and lower backslope-footslope-toeslopes while shoulder-upper backslope soils are yellow-brown earths. *Chionochloa crassiuscula* is prominent in the vegetation on all sites. As nothing suitable was found for ^{14}C dating these soils, a precise estimate of their age is not possible. They are the most developed soils seen in Noisy Creek and are similar to the 7-10,000 year soils at Hut Flat.

The glacially scoured bedrock surfaces at the head of Noisy Creek provide a contrast with those bedrock surfaces near the basin outlet. At the valley head a broad U-shaped valley cut in bedrock is dissected by several streams incised in narrow, vertical sided ravines more than 50 m deep (Fig. 49). The interfluve areas are mainly moderately sloping (10-20°) and appear to be relatively unmodified, glacially scoured bedrock. All soils in this area were recent or weakly developed yellow-brown earths. Turf banked terraces with recent soils (Fig. 31) were common. Nothing was found for ^{14}C dating, hence it is impossible to be sure whether these surfaces have been deglaciated recently (<1000 years), or continual mass movement since deglaciation in the early Aranuian has maintained young soils at the surface.

5.6 TOP CROPP CIRQUE

Geomorphology and soils of the cirque at the head of Cropp River (Fig. 53) indicate recent deglaciation, with active mass movement since deglaciation. A radiocarbon date of 2370 $\pm$ 70 years B.P. (NZ 6317) was obtained from the base of a peat (about 90 cm below the surface), overlying bedrock, near the edge of the cirque. This provides a minimum estimate for deglaciation. The U-shaped glacial valley form has been weakly modified by erosion and deposition, with glacial striations still present on some exposed bedrock surfaces.

Only two landforms occur in the top Cropp cirque. Steep (40-90°) rock hillslopes form the cirque perimeter. These are the sediment sources for the depositional features and are commonly near vertical, with no soil or vegetation cover. Debris cones at the base of the rock slopes are moderately to steeply sloping (15-40°), and have formed by rockfall, debris and snow avalanching and fluvial deposition from the rock slopes. Large boulders are common on the surface of the cones. Colluvium is up to 20 m thick in the cirque floor and decreases in thickness towards the cirque margins. Colluvium is thicker on the west side of the cirque than on the east. Streams are deeply incised in the debris cones, forming narrow bedrock-floored channels with vertical sides.

The soil pattern comprises mainly recent soils, with minor areas of older soils. The most developed soil observed in free draining sites was a weakly podzolised and gleyed soil similar to F3 ($A_h/E_{agj}/B_w/C$ horizon sequence). These soils

R.S. Rock slopes
D.C. Debris cones
Radiocarbon dating site
1 NZ 6317(2370 ± 70 yrs BP)

Fig. 53: Top Cropp cirque.

occur in the distal parts of debris cones and near the edge of the cirque. Recent soils dominate towards the head of the cirque and over most of the debris cones. Towards the upper margins of the debris cones recent soils with indistinct sedimentary layering are common, resulting from frequent mass movement on the very steep hillslopes above. Abundant evidence of small scale mass movements was seen. Organic soils are also common on the debris cones since much runoff is generated from the bare rock slopes above.

5.7 CONCLUSIONS

The three main study areas (mid-Cropp, Noisy Creek, and top Cropp) provide interesting comparisons in geomorphology and soil patterns. These will be briefly considered here, but further discussion of the significance of radiocarbon dates and observed soil patterns is contained in Chapter 9 (soil erosion history) following consideration of soil analytical data and soil-vegetation relationships (Chapters 6 and 7).

The top Cropp cirque has only recently been deglaciated and the glacial cirque is weakly modified by steeply sloping debris cones in footslopes. The soil pattern is dominated by recent soils and, in the upper parts of debris cones, by recent profiles with indistinct sedimentary layering. Soils reflect the young age of the landscape and active geomorphic processes.

In the mid-Cropp region, there is a partial record of landscape evolution over at least 10,000 years. Till dated at 10,250 $\pm$ 150 years B.P. (NZ 6576) establishes the presence of ice near Reckless Torrent at that time, and provides a likely maximum age for much of the upvalley landscape. Glacial landforms have largely been modified and are discontinuous in their distribution. The oldest landscape components recognised were:

- 10,250 year old moraine near Reckless Torrent;
- 7000 year old alluvial cone surfaces at Hut Flat (i.e. T4 and equivalents);

- stable ridge crests at Hut Flat (e.g. site of S6);
- older debris cone surfaces on the south bank of Cropp River below Hut Creek.

In area, these are minor landscape components. No surfaces in the age range from 2000 to 7000 years B.P. were recognised. This may be because it was a period of erosional removal of valley fill sediments, or alternatively landforms constructed at this time were removed in later episodes of erosion (i.e. erosion censoring of the record). Dates from Cropp riverbed at Hut Flat indicate that the river was at a lower level in the period 1700-3200 years B.P. (NZ 6113, 6114) and has subsequently aggraded considerably. Thus land surfaces formed between 3200 and 7000 years B.P. may have been destroyed. Events younger than 1500 years B.P. are well recorded in the array of young terraces (T1, T2, T3) and associated stratigraphy. On steep slopes there are scattered remnants of older slope facets that maybe correlated with the older depositional landforms. However the steep slopes are dominated by recent soils and weakly developed yellow-brown earths as a result of active erosion processes.

The dates from Hut Flat also indicate that the gross landscape features were present by 7000 years B.P. The exit to Hut Flat has remained relatively stable, due to the presence of bedrock resistant to fluvial incision, and the delivery of coarse sediment from Hut Creek. Base level lowering has been limited to perhaps 20 m (estimated from discordance of slopes adjacent to exit from Hut Flat, and the level to which the surface of T4 is graded). The changes in base level at Hut Flat and Noisy Creek are comparable.

Noisy Creek has been subject to little base level lowering and at first appearance seemed to be a stable area of considerably antiquity. A high degree of preservation of glacial features and post-glacial debris cones and alluvial terraces, plus the presence of a lake, suggested that this would be an area of relative stability. Although a radiocarbon date of 7780 $\pm$ 110 years B.P. (NZ 6348) indicated that this area had been deglaciated for a considerable length of time, few old soils remain anywhere in Noisy Creek. Studies at Noisy Creek showed that Pleistocene and Aranuian glacial advances formed a hollow behind the Pounamu ultramafics prior to 8000 years B.P. During deglaciation a lake was present in this hollow and debris cones formed that were graded to the lake as base level. This occurred during the period >8000 to c. 4000 years B.P. Incision of Noisy Creek at the lake outlet (after 4000 years B.P.) led to lowering of lake level and formation of benches along the truncated distal margins of debris cones and in bedrock. Eventually the lake was infilled and the present flood plain formed. Most lake sediments above the present basin outlet were removed during this phase.

The lack of old soils at Noisy Creek results from two causes. In the valley bottom all terraces post-date the relatively recent infilling of a lake. Recent soils predominate except on some of the higher terraces where weakly gleyed and podzolised soils occur. On hillslopes and debris cones the dominance of recent soils is a result of frequent mass movements and fluvial erosion/deposition. Despite the apparent stability implied by a relatively stable base level, geomorphic processes have been extremely active and have resulted in a dominance of young soils throughout Noisy Creek.

CHAPTER 6 - CHEMICAL AND PHYSICAL CHARACTERISTICS OF SELECTED SOIL PROFILES

6.1 INTRODUCTION

Chemical weathering occurs because rocks and minerals are seldom in equilibrium with near surface waters, temperatures and pressures (Birkeland, 1974). Elements released by weathering may be leached, adsorbed, reprecipitated, or assimilated by plants and organisms, depending on the nature of the soil environment. The behaviour of the various elements depends on their individual physico-chemical properties (particularly solubility in aqueous systems and interactions with other soil components) and the leaching potential of the soil solution (the net result of the interaction of soil water balance and interflow characteristics - Young *et al*, 1977).

Weathering of siliceous sediments under conditions of continued leaching and unimpeded drainage results in loss of monovalent and divalent cations (and balancing anions), and the residual accumulation of Fe, Al and Si (Chesworth, 1973b). The relative mobility of Fe, Al and Si is controlled by pH, Eh and chelation. Over normal soil pH range (5-7) Fe and Al are less soluble than Si (laterization). With decreasing pH and/or chelation, Fe and Al become more soluble than Si (podzolisation). In reducing environments the mobility of Fe (as Fe^{2+}) is further enhanced. Polynov (1937), Crompton (1960) and Glazovskaya (1968) describe the relative mobility of the major elements for normal, acid chelating and reducing environments (Table 7). Previous studies in Westland environments of high leaching potential (Stevens, 1968; Campbell, 1975; Ross *et al*, 1977; McKie, 1978) show that the most mobile elements (Na, Mg, K, Ca) are commonly removed entirely from the soil system, and that the relative mobility of Fe, Al and Si is controlled by pH and Eh.

In view of the high rainfall of Cropp basin, this study concentrated on analysis of the least mobile elements in the

Mobility phase	Polynov (1937)	Crompton (1960), Glazovskaya (1968) acid chelating	reducing
Phase I	Cl^{-}, SO_4^{2-}	Cl^{-}, SO_4^{2-}	Cl^{-}, SO_4^{2-}
Phase II	Ca^{2+}, Na^{+}, Mg^{2+}, K^{+}	Ca^{2+}, Na^{+}, Mg^{2+}, K^{+}, Fe^{3+}, Al^{3+}	Ca^{2+}, Na^{+}, Mg^{2+}, K^{+}, Fe^{2+}, Mn^{2+}
Phase III	SiO_2, P	SiO_2, P	SiO_2, P
Phase IV	Fe_2O_3, Al_2O_3	-	Al_2O_3

TABLE 7: Element mobility series.

soil system. Emphasis in this chapter is on relating the analytical data to the ages of the sampled soils. Discussion of soil processes and comparison with other studies is included in Chapter 8.

6.2 OBJECTIVES

1) To characterise changes in chemical properties with soil age to support the interpretation of soil morphology and stratigraphy.
2) To characterise a range of soils presently mapped within the McKerrow soil set.
3) To characterise the nutrient status of selected soils in Cropp drainage basin.

6.3 SELECTION OF SOILS FOR ANALYSIS

Analysis has concentrated on the soils of the two sequences described in chapter 5 i.e. F1 to F7, and S1 to S6. In addition the 3 recent soils (typical of those that form the land surface over most of Cropp basin; S7, S8, S9) and one old, shallow soil over bedrock (S10) were analysed.

6.4 SAMPLING PROCEDURES

All soils were sampled from one side of the pits excavated for detailed profile description. Samples of each soil horizon were removed and stored in polythene bags. Horizons greater than 10 cm in thickness were subsampled. Soils were mostly sampled on a non volume-weight basis, although a number of samples (representative of the range of bulk density values) were taken with a 10 cm diameter bulk density corer. Because of the stony nature of many of the soils, and the abundance of large roots in A_h horizons, these samples are of dubious accuracy for quantitative analysis. Because of the practical difficulty of obtaining adequate volume-weight samples,results of soil chemical analysis are expressed as grams per 100 grams of soil fine-earth fraction (% by weight).

6.5 ANALYTICAL PROCEDURES

6.5.1 Sample preparation

Half of each sample was air-dried in the laboratory, crushed gently with a rolling pin and sieved through a 2 mm sieve to remove stones. The <2 mm fraction was used for all chemical analyses. Part of this fraction was ground using a pestle and mortar to pass a 250 µm sieve. This fraction was used in the determination of total C and N, 0.5M H_2SO_4 soluble-P, organic P and dithionite-citrate-extractable Fe and Al. Another fraction was ground using a tungsten carbide vessel in a Tema mill to pass a 150 µm sieve. This fraction was used for total element analysis.

6.5.2 Choice of analyses

Parameters already proven to be good indicators of pedogenic status in environments with high leaching potential were chosen for analysis (Stevens, 1968; Campbell, 1975; McKie, 1978). The following analyses were carried out on all soils: pH (in water and NaF); acid ammonium oxalate-extractable Fe, Al, and Si; total element analysis (for Ca, P, Fe, Al, Si, Zr); loss-on-ignition; and particle size analysis. Total element analysis was chosen in preference to detailed P fractionation, which has been widely used to investigate soil development (Chang and Jackson, 1957; Stevens, 1968; Williams *et al*, 1969; Syers and Walker, 1969a, b; Smeck, 1973; Adams and Walker, 1975; Walker and Syers, 1976). Total element concentrations could be determined rapidly by X-ray fluorescence spectroscopy, whereas P fractionation is a long, complex procedure that was impractical with such a large number of samples to be analysed (approximately 200). Major changes in total element composition with soil development were anticipated in this environment.

For a few samples other parameters were chosen primarily to indicate nutrient status. These included: exchange properties (exchangeable Ca, Mg, K, Na; cation exchange capacity; KCl-extractable Al and H), acid-soluble and organic phosphorus, organic carbon, total nitrogen; also pyrophosphate- and dithionite-citrate extractable Fe and Al (for comparison with oxalate-extractable Fe and Al).

6.5.3 Analytical techniques

With the exception of total element and particle size analysis, all analyses were replicated.

(a) pH (on <2 mm, air-dried soil).

Soil: distilled water suspensions (1:2.5 for most samples, 1:5 for peaty samples) were stirred vigorously, left to stand overnight, and after further stirring read on a Radiometer 23 pH meter equipped with glass-calomel electrodes. The pH of soil: 1M KCl suspensions was determined similarly, for some samples.

pH in NaF was measured using a sample : 0.85 M NaF ratio of 1:50 following Campbell's (1975) modification of the Fieldes and Perrott (1966) test for active Al. The soil/NaF suspension was stirred vigorously for 1 minute and read after a further minute. The pH of the 0.85 M NaF solution was checked prior to use, and if outside the range 7.0 to 8.0 it was discarded (Blakemore *et al*, 1981).

(b) Oxalate-extractable Fe, Al and Si.

A modified approach to that of McKeague and Day (1966) was used. 1 g of soil (<2 mm, air-dried) and 50 ml of 0.2M ammonium oxalate (pH 3.0) were shaken for four hours in darkness, on an end-over-end shaker. After addition of 'Superfloc' and centrifuging, aliquots were prepared for analysis:

- Fe by atomic absorption and Al by atomic emission
- Si was determined colorimetrically following the method of Weaver *et al* (1968).

(c) Pyrophosphate-extractable Fe and Al.

1 g soil (<2 mm, air-dried) and 100 ml 0.1M sodium pyrophosphate were shaken overnight (16 h) on an end-over-end shaker (McKeague, 1967). After addition of

'Superfloc', ultra centrifuging and suitable dilution, Fe and Al were measured by atomic absorption.

(d) Dithionite-citrate extractable Fe and Al.
1 g soil (<0.25 mm, air-dried), 1 g sodium dithionite and 50 ml 22% sodium citrate solution were shaken overnight (16 h) on an end-over-end shaker (Holmgren, 1967). After the addition of 50 ml of distilled water and 'Superfloc', the mixture was shaken vigorously, centrifuged, and filtered. An aliquot was treated with acid and digested on a water bath for 1 hour. After cooling Fe and Al were measured by atomic absorption.

(e) Total element analysis.
Total element concentrations were determined by X-ray fluorescence spectroscopy (XRF) on soils ground to pass 150 µm. Major elements (Ca, P, Fe, Al, Si) were measured on fusion beads prepared following the general methods of Norrish and Hutton (1969), with modifications after Harvey *et al* (1973) and Schroeder *et al* (1980). Samples were fused at 1000^{o} with a lithium tetraborate/lithium carbonate/lathanum oxide flux, which eliminates grain size effects and diminishes absorption and enhancement effects during sample measurement. Zirconium was measured directly on pressed powder discs. Sample measurement was undertaken on an automated Philips PW 1400 Spectrometer in the X-ray Spectrometry Laboratory, Department of Geology, University of Canterbury. Standards used to check instrument calibration were geological in origin. As the samples from this study were the first soils to be analysed by this laboratory a few chemically analysed samples from the Reefton chronosequence (Campbell, 1975), covering a range of organic matter contents, were analysed as a check that the results obtained were reasonable. Duplicate beads, made for a few Cropp samples, and duplicate analysis of a number of beads indicated that the precision of the method was extremely good.

(f) Loss on ignition (LOI).
This was determined by igniting 5 to 10 g samples of soil (<2 mm, oven dried) at 550°C. The duration of ignition varied, depending on the organic content of the sample, but was generally 2-3 hours.

(g) 0.5M H_2SO_4 - soluble P.
0.5 g of soil (<0.25 mm, air-dried) and 100 ml 0.5M H_2SO_4 were shaken overnight (16 h) on an end-over-end shaker. After filtering, an aliquot was taken and P measured colorimetrically (Blakemore *et al*, 1981).

(h) Organic P.
0.5 g of soil (<0.25 mm, air-dried) was ignited, to convert organic P to acid-extractable P, and then extracted and measured as for 0.5M H_2SO_4 - soluble P.

(i) Exchange properties.
Samples of air-dried, <2 mm soil were leached with 1M ammonium acetate (pH 7.0), washed with ethanol, leached with 1M NaCl, distilled, and cation exchange capacity determined on the ammonia absorbed in boric acid (Blakemore *et al*, 1981). Exchangeable Ca, Mg, K and Na were determined on the ammonium acetate leachate by atomic absorption or emission. Exchangeable Al and H were determined separately by extracting 10 g of soil (<2 mm, air dry) with 50 ml 1M KCl for 16 hours. The titration procedure of Yuan (1959) was used to measure Al and H.

(j) Organic C.
This was determined on <0.25 mm, air-dried soil using the wet dichromate oxidation of Walkley and Black (1934), as modified by Walkley (1947), and titration against ammonium ferrous sulphate. This method recovers 70-75% of the organic carbon present, so results were multiplied by 1.3 to estimate total organic carbon.

(k) Total N.
This was determined on <0.25 mm, air-dried soil using the

semi-micro Kjeldahl method to convert nitrogen to ammonium, which was then determined by the distillation and titration procedure of Blakemore *et al* (1981).

(1) Particle size analysis.

Air-dried samples were sieved through screens with square openings of 50.8, 19, 12.7, 6.35 and 2 mm and the weight of rock fragments recorded. It should be noted that large clasts (>50 mm) were under-sampled in the field because of practical limitations to the size of sample that could be collected and transported back to the laboratory. It was impractical to sieve samples in the field. Additionally many clasts broke up during sieving, thus altering the natural particle size distribution. These limit the interpretation of the particle size analyses.

The <2 mm fraction was lightly rolled to crush aggregates, and quartered to give subsamples of 10-20 g for detailed analysis. Subsamples were treated with boiling H_2O_2 to remove organic matter. Samples that gave a positive reaction to the phenolpthalein test (Fieldes and Perrott, 1966) were treated with dithionite-citrate (Mehra and Jackson, 1960). Samples were then dispersed in 'Calgon' (66 g sodium hexa metaphosphate + 14 g sodium carbonate per litre). The treated and dispersed samples were washed through a 63 µm sieve. After oven drying, the >63 µm fraction was sieved at half ϕ intervals. The <63 µm fraction was analysed using a Micrometrics Sedigraph 5000 particle size analyser.

International size classifications were adopted: sand = 2 - 0.02 mm; silt = 0.02 - 0.002 mm; clay = <0.002 mm.

6.6 RESULTS AND DISCUSSION

Analytical data for the two soil sequences (F1 to F7, S1 to S6) are tabulated in Appendix 3. Data for S7, 8, 9, 10, exchange chemistry, phosphorus, organic matter, pyrophosphate- and dithionite-citrate extractable Fe and Al is tabulated in the text at the point each set is discussed. For comparison and interpretation data are presented diagrammatically in this section as soil chemical property versus soil depth plots. All depth measurements are taken from the soil surface and include organic horizons. Emphasis is given to changes in the upper 50 cm of the soil profile.

The use of concentrations by weight indicates relative losses, gains and redistribution of elements during soil development. The absolute losses and gains have not been estimated because of the impracticality of volume-weight sampling. In the accompanying figures and tables total element analyses, oxalate-extractable Fe, Al and Si, pyrophosphate- and dithionite-citrate extractable Fe and Al are expressed as proportions of the ignited weight of the sample, to remove the diluting effect of organic matter in surface horizons (Blume and Schwertmann, 1969). Particle size analysis, cation exchange data, acid-soluble and organic phosphorus, organic carbon and total nitrogen analyses are expressed on an oven-dry basis.

Data are discussed in terms of the two sequences of soils described in Chapter 5 (F1 to F7, S1 to S6) and the other commonly occurring soils (S7 to S10) - see Figs 44 and 45. For soils on gentle slopes, F1 to F5 are members of a post-incisive, non-strict chronosequence. Trends in the data are most regular for this group of soils. F6 (7,000 to 10,000 years; formed in bedrock) and F7 (2,500 years; formed in fine textured alluvium) are not directly comparable with F1 to F5 due to their different parent materials. Trends in the data for the sequence from steep slopes are less regular. S1, S2, $S3_u$, S4, $S5_u$, and S6 are a post-incisive, non-strict chronosequence also, but with some variation in site characteristics and parent materials. These soils are all in well drained sites, while sites for $S3_l$ and $S5_l$ are less well drained.

6.6.1 Soil pH (H_2O)

a) soils on gentle slopes.

Surface horizon pH decreased progressively from 5.1 in F1 to 4.0 in F6 (Fig. 54). In all soils, pH increased with depth, and most C horizon values were in the range 5.4 to 5.8. The C horizon values were far lower than those obtained from rock samples and riverbed alluvium. Ground rock samples gave pH's ranging from 8.8 for a metapsammite to 7.3 for a metapelite. Alluvium (<2 mm) from Cropp riverbed ranged between 7.8 and 8.2. This suggests that the designated C horizons were also weathered.

F7 (2,500 year old soil in fine textured alluvium) had a lower surface horizon pH than F4 (the 7,000 year old soil). The surface organic horizon of the peaty soil F4p had lower pH than the equivalent horizon of the adjacent mineral soil, F4, and low pH persisted throughout the organic horizons of F4p. The lower pH in $F4_p$ reflects both increased runoff in topographically lower areas causing increased leaching, and the dominance of organic acids in the peaty surface horizon.

b) soils on steep slopes.

Soils on steep slopes tended to have lower pH throughout the profile compared to soils on gentle slopes, reflecting differences in vegetation and parent material. Variability in parent material was greater on steep slopes, where soils were formed in pre-weathered colluvium. Debris avalanches on steep hillslopes commonly left a veneer of colluvium over bedrock. Samples from the surface of two recent (<2 years) debris avalanches had pH values of 5.0 and 5.2.

There was a general trend in this sequence for a decrease in pH of surface horizons with age, from 4.8 in S1 to 3.8 in S6 (Fig. 55). Comparison of surface values was complicated by variation in organic matter content (measured by LOI) from 18-80%. Wherever LOI was high then pH was low e.g. S2 (140 years) had a LOI of 45% and surface horizon pH of 4.2, whereas $S3_u$ (500 to 1,000 years) had a LOI of 20% and a pH of 4.4. In all soils, pH increased with depth and most C horizons ranged from 4.9 to 5.4. $S3_l$ had significantly higher subsoil values

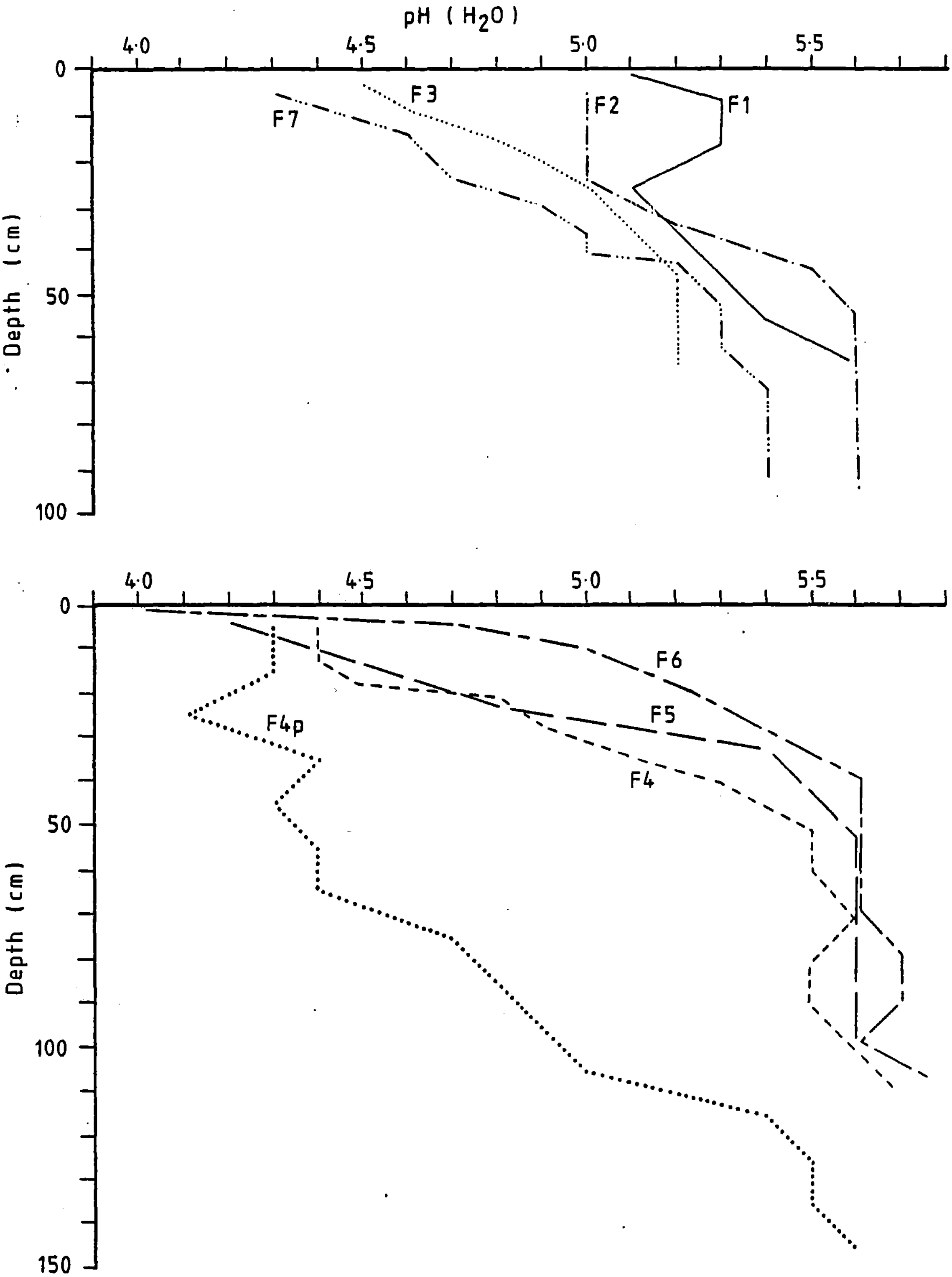


Fig. 54: pH (H_2O) - soils on gentle slopes.

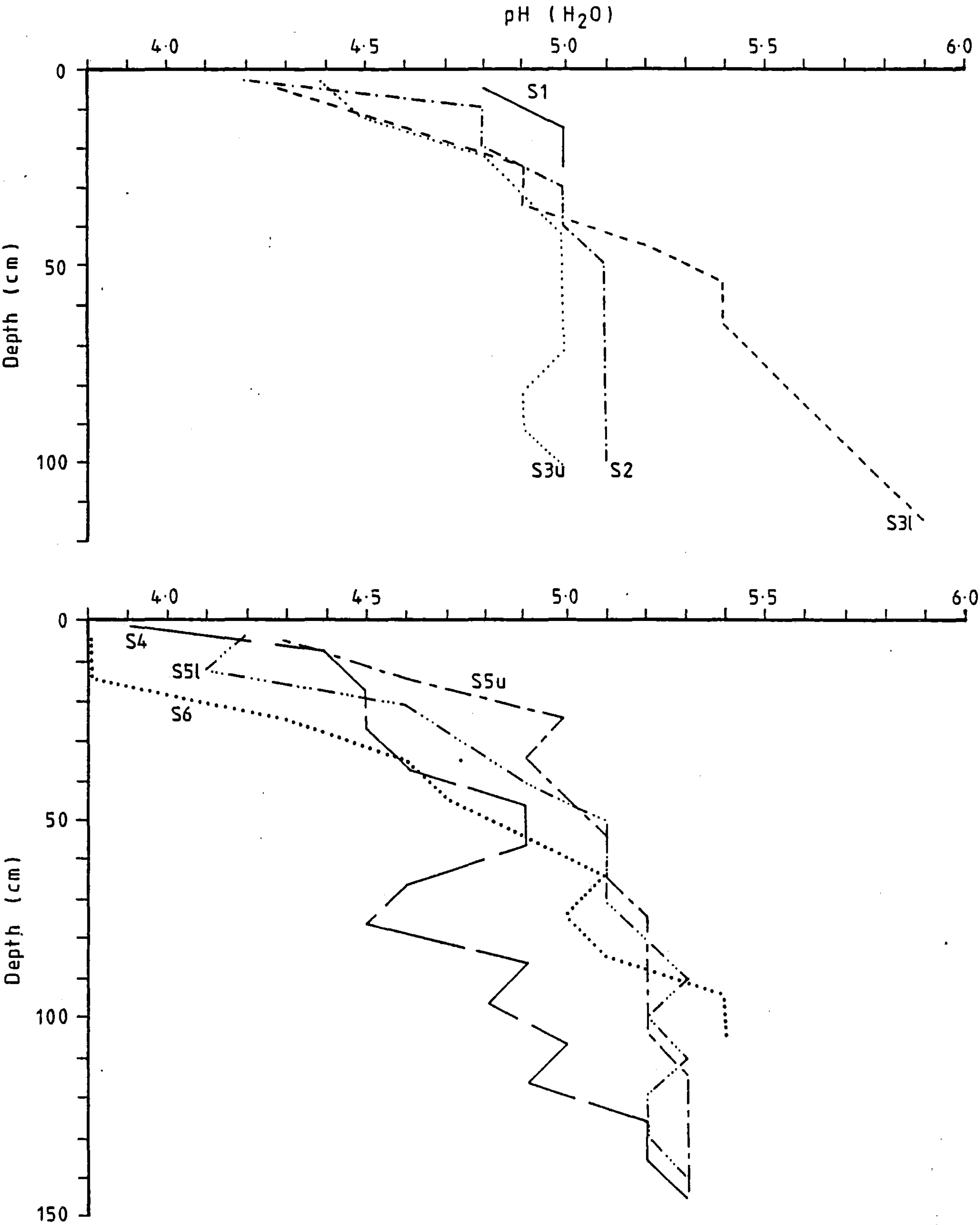


Fig. 55: pH (H_2O) - soils on steep slopes.

(5.6 to 5.9) while S4 had a variable subsoil pH. The two pairs of soils on risers ($S3_{u,l}$, $S5_{u,l}$) both had a lower surface horizon pH in the soil in the lower slope position. The same trend was noted for soils on gentle slopes ($F4/F4_p$).

6.6.2 pH (KCl)

Soil pH measured in electrolyte solutions gives more reproducible results, as they are not influenced by the salt content of the soil, and the values obtained depend less on the electrode position in the suspension (Blakemore *et al*, 1981). 1M KCl and 0.01M $CaCl_2$ are the most commonly used salts. When 1M KCl is used, extensive ion exchange takes place, including the release of Al^{3+} and H_3O^+. Values of pH obtained in 1M KCl are usually lower than those obtained in water. Mekaru and Uehara (1972) describe how the ΔpH value (pH (1M KCl) - pH (H_2O)) can be used to indicate whether soils with variable charge have a net positive or negative charge.

pH (KCl) was measured only for six of the soils on gently sloping sites (F1, F2, F4, F4p, F5, F6). The values obtained were all lower than pH (H_2O) - Appendix 3. In general terms pH (KCl) shows similar trends to pH (H_2O), namely a decrease in surface horizon pH (KCl) with increasing age, and an increase in pH (KCl) down the profile. F1 and F2 however, show a decrease in pH (KCl) with depth in the profile.

ΔpH values (Table 8) were all negative and became increasingly negative with depth in the profile. Values of ΔpH for surface horizons tended to become less negative with increasing soil age (-1.5 in F1 to -0.8 in F6). Negative values of ΔpH indicate that these soils carry a net negative charge.

6.6.3 pH (NaF)

Fieldes and Perrott (1966) used pH measured in a soil: NaF suspension as an indication of allophane (poorly ordered hydroxy aluminosilicates) in soils. Aqueous solutions of fluorides at pH >7 displace hydroxyl ions from hydroxyalumina sites on allophanes, causing a rise in pH of the solution.

Sample Code	F1	F2	F3	F4	F4p	F5	F6
.1	-1.5	-1.2	n.d.	-1.1	-0.7	-0.9	-0.8
.2	-1.9	-1.2	n.d.	-1.2	-0.8	-1.2	-1.5
.3	-2.1	-1.2	n.d.	-1.1	-0.5	-1.2	-1.6
.4	-1.8	-1.8	n.d.	-1.1	-0.8	n.d.	-1.7
.5	-1.9	-2.2	n.d.	-1.3	-0.7	-1.1	-1.6
.6	-2.0	-2.1	n.d.	-2.0	-0.7	-1.2	-1.8
.7	-2.1	-2.1	n.d.	-2.0	-0.7	-1.5	-1.5
.8	-2.4	-2.1	n.d.	-1.9	-0.9	-1.4	-1.7
.9		-2.3	n.d	-2.2	-0.9	-1.2	-1.6
.10		-2.2		-1.7	-0.7	-1.4	-1.5
.11				-1.6	-0.8	-1.4	-1.4
.12				-2.0	-1.3	-1.2	-1.4
.13				-2.1	-1.3		-1.9
.14					-1.6		
.15					-1.7		

n.d. not determined

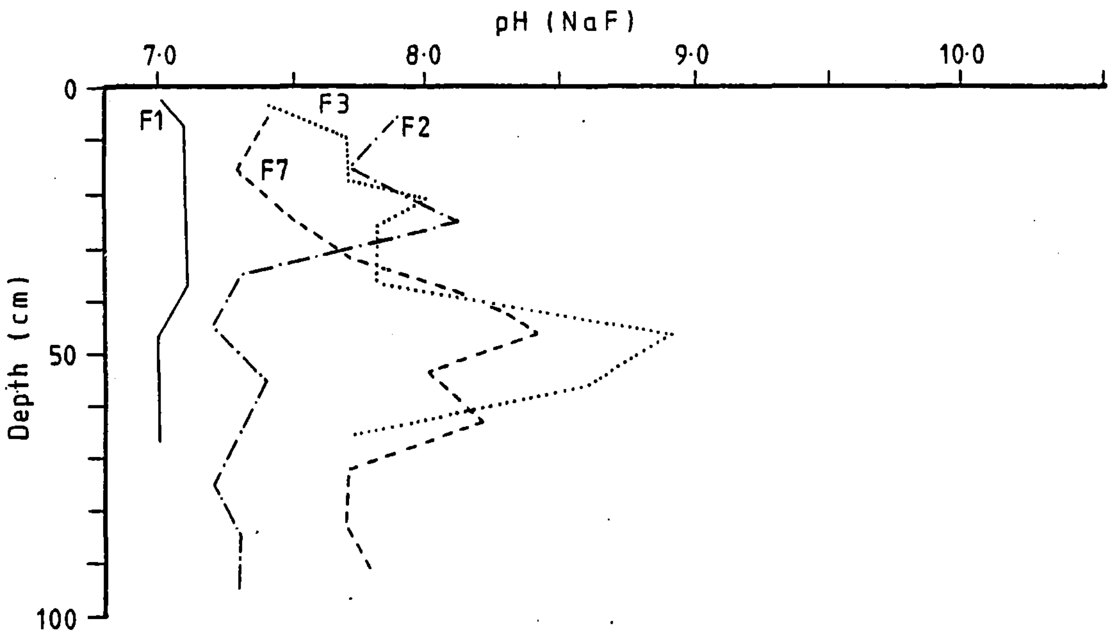
TABLE 8: ΔpH values for soils on gentle slopes.

The test is not specific to allophane, but yields positive results for a range of materials with reactive hydroxyalumina sites (Bracewell *et al*, 1970; Brydon and Day, 1970; Perrott *et al*, 1976). A careful choice of starting pH and reaction time does reduce the reaction due to other phases (Brown *et al*, 1978). The method adopted in this study, which ensured the pH of 0.85 M NaF solution was between 7 and 8, and measured the pH attained after 2 minutes, is essentially that used by N.Z. Soil Bureau (Blakemore *et al*, 1981).

pH (NaF) is well correlated with oxalate-extractable Al (Brydon and Day, 1970; Campbell, 1975; Harrison, 1982) and provides an index of oxalate chemistry. High pH (NaF) values are found in illuvial horizons of podzolised soils where amorphous hydroxy aluminosilicates accumulate. The significance of this is discussed further with reference to oxalate chemistry.

a) Soils on gentle slopes (Fig. 56)
F1 had uniformly low values of pH (NaF) (7.0 to 7.1) while F2 showed a decrease in pH (NaF) with depth, from 7.9 to 7.3. In F3 to F5 maximum values of pH (NaF) occured in the B horizon and increased from 8.0 in F3, to 8.8 in F4, to 10.2 in F5. F7 had a B horizon value of 8.4. Some C horizon values were also relatively high (8-9). The general trend in the older soils was for low values in the A and E horizons and maximum values in the illuvial horizons (B_w, B_s, B_f), indicative of podzolic soils. The maximum value recorded was 10.2, from the B_f horizon of F5, but most samples gave values less than 9.0.

b) Soils on steep slopes (Fig. 57)
There is a change from low values (7.0 to 7.5) and a slight increase with depth (up to 8.5) in the younger soils (S1, S2, S3) to a trend of lower surface horizon values (<7.0) and higher subsoil values (9.5 to 10.0) in the older soils (S4, $S5_u$, $S5_l$, S6). The highest value (10.1) was obtained from the base of the B_s horizon of S6.



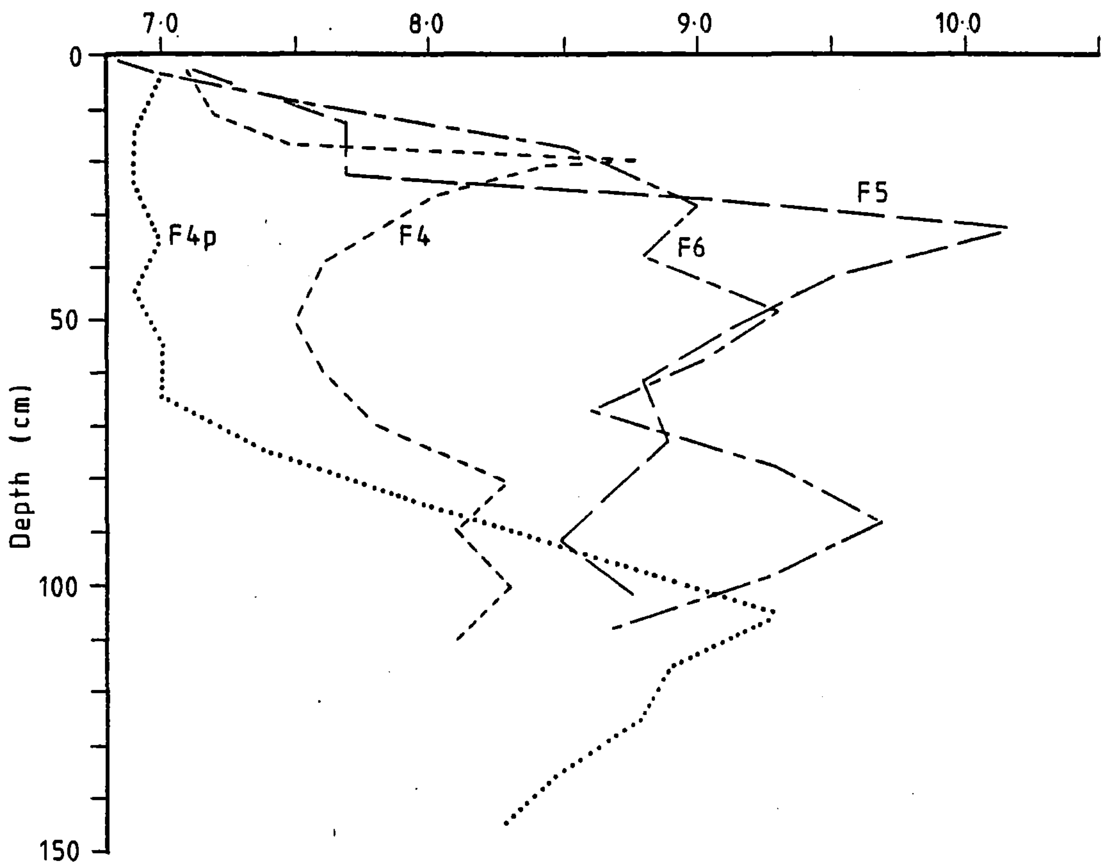


Fig. 56: pH (NaF) - soils on gentle slopes.

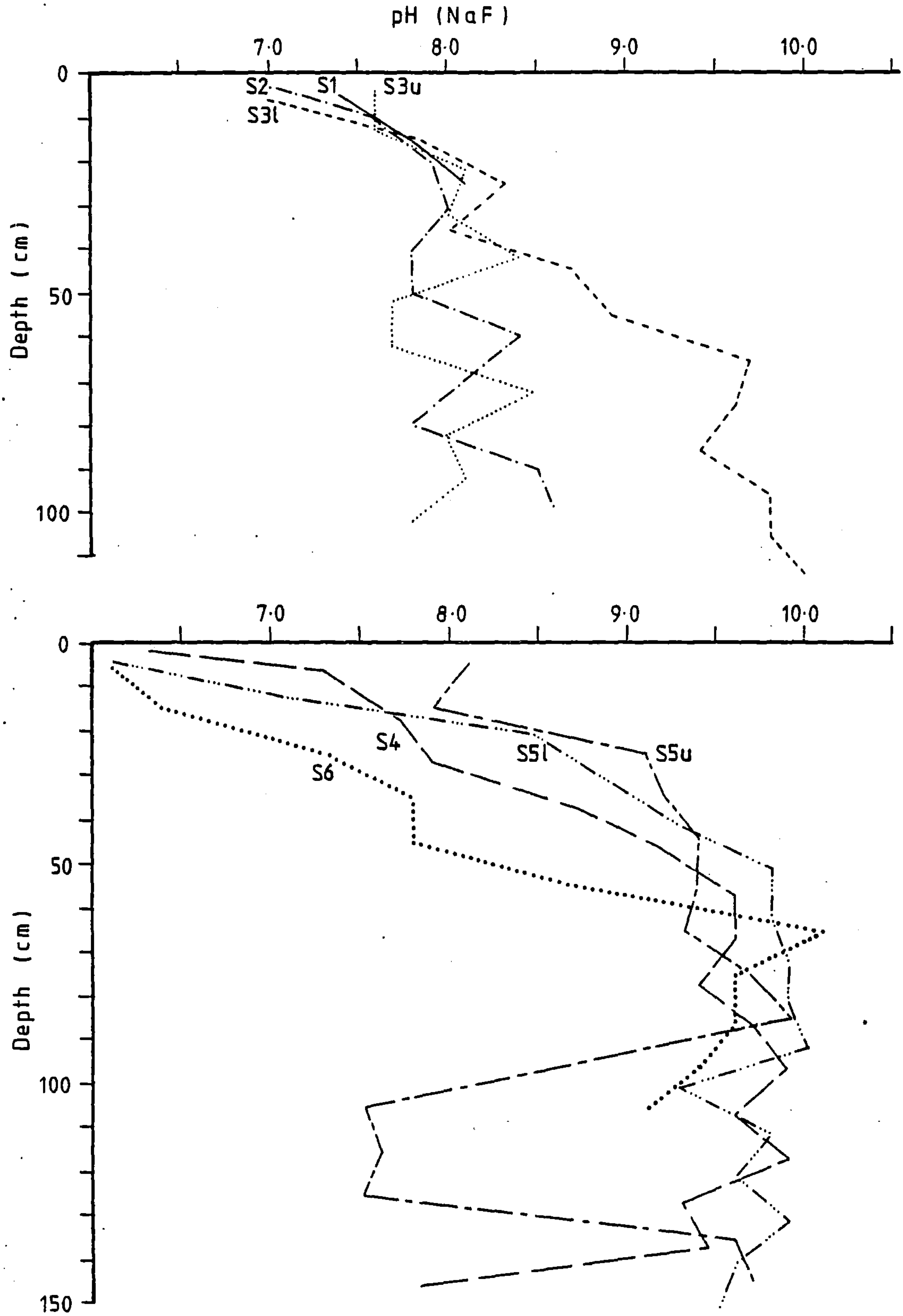


Fig. 57: pH (NaF) - soils on steep slopes.

In detail, trends with age are not as clear as for soils on gently sloping sites. $S3_l$ had markedly higher C horizon values than $S3_u$ (although both are the same age) while S4 (1,000 years) had similar values to the 7,000 to 10,000 year old soils ($S5_u$, $S5_l$, S6). The organic surface horizons of S4, $S5_l$, and S6 had very low (<7.0) values of pH (NaF).

6.6.4 Extractable iron, aluminium and silicon

The oxides of Fe and Al are important constituents of soil clays that occur in several mineralogical forms (crystalline, X-ray amorphous, complexes with organic matter). The amounts and distribution of the various forms indicate the intensity and direction of soil formation. Many reagents have been used to extract the various forms of Fe and Al, and associated Si, in soils, in order to investigate soil forming processes and changes with time. The most widely used selective dissolution extractants are:

(a) acid ammonium oxalate (simply referred to as oxalate hereafter) - this reagent is the most useful single extractant for amorphous inorganic forms of Fe, Al and Si (Schwertmann, 1959, 1964; McKeague and Day, 1966; McKeague, 1967; Blume and Schwertmann, 1969; McKeague *et al*, 1971; Farmer *et al*, 1983). Oxalate also extracts Fe and Al from organic complexes (McKeague, 1967; Farmer *et al*, 1983) but this can be estimated separately using pyrophosphate. While crystalline oxides are partially dissolved by oxalate (McKeague *et al*, 1971; Arshad *et al*, 1972; Pawluk, 1972; Schwertmann, 1973; Walker, 1983) the method is widely used to give a measure of the quantity of amorphous oxides in soils.

(b) 0.1 M sodium pyrophosphate (pyrophosphate) - this reagent selectively extracts organic complexes of Fe and Al and minor amounts of amorphous inorganic forms (McKeague, 1967; Bascomb, 1968; McKeague *et al*, 1971; Farmer *et al*, 1983).

(c) dithionite-citrate (dithionite) - this reagent extracts all free (non-silicate) Fe oxides including crystalline, amorphous and organically bound Fe, and partially extracts free Al oxides (McKeague and Day, 1966; McKeague *et al*, 1971; Blume and Schwertmann, 1969; Farmer *et al*, 1983).

Table 9 summarises the reactions of these three reagents with the various forms of Fe, Al, and Si in soils. These selective dissolution techniques can be employed to determine the total free iron oxides (dithionite-extractable), amorphous inorganic iron oxides (oxalate-extractable minus pyrophosphate-extractable), and organic complexed iron oxides (pyrophosphate-extractable). The differential extraction of aluminium oxides with these three reagents is less well extablished (McKeague *et al*, 1971). These selective dissolution techniques are empirical methods that cannot differentiate sharply between different forms of Fe and Al, which are present as various continua in natural soil systems (Schwertmann, 1973). The amount and distribution of different forms of extractable Fe and Al indicates the intensity and conditions of soil formation.

Selective dissolution techniques have been widely used to evaluate relative soil development. The absolute amounts of the different forms of extractable Fe and Al, or ratios of the various forms, have both proved useful (Saunders, 1965; Blume and Schwertmann, 1969; Alexander, 1974; Campbell, 1975; Ross *et al*, 1977; Harrison, 1982; Alexander and Holawaychuk, 1983; Arduino *et al*, 1984; Tonkin, 1984; McFadden and Hendricks, 1985; Rebertus and Buol, 1985). Studies of soil development in Westland (Mokma *et al*, 1973; Campbell, 1975; Ross *et al*, 1977; Smith and Lee, 1984) have shown that oxalate-extractable Fe and Al increase with increasing soil development, until pH (H_2O) falls below 4.5. Thereafter topsoil amounts decline, and B horizon amounts increase. This characteristic pattern results from dissolution and mobilisation of Fe and Al as low pH develops in the A and/or E horizons with time, combined with translocation of Fe and Al

Element in specified component	0.1M pyro-phosphate	Acid ammonium oxalate	Dithionite-citrate
Al in			
Humus	strong	strong	strong
Hydrous oxides			
non-crystalline	weak	strong	moderate
crystalline (gibbsite)	none	none	weak
Fe in			
Humus	strong	strong	strong
Hydrous oxides			
ferrihydrite	weak	strong	strong
goethite, hematite, etc	none	none	strong
Si in			
Opaline silica	none	none	none
Crystalline silica	none	none	none
Si and Al (Fe)			
Allophane	weak	strong	moderate
Imogolite	weak	strong-moderate	weak
Layer silicates	none	none	weak

TABLE 9: Dissolution of Fe, Al and Si in various soil constituents with different analytical reagents (from Parfitt, 1980).

as electropositive sols, or chelates, and subsequent deposition in the B horizon where pH is above 5 (discussion of modern concepts of podzolisation is contained in Farmer *et al*, 1980; Flach *et al*, 1980; Farmer and Fraser, 1982; Anderson *et al*, 1982; Childs *et al*, 1983; Buurman and van Reeuwijk, 1984). Furthermore, studies of soil development in Westland show that as soil development proceeds towards gley podzols, amounts of oxalate-extractable Fe and Al decrease as poorly ordered materials are dissolved, or converted to crystalline clay minerals and hydrous oxides (Mokma *et al*, 1973; Campbell, 1975; Ross *et al*, 1977; Mew and Lee, 1981).

Ratios of the various forms of extractable Fe and Al have also been shown to vary with age. The ratio of oxalate-extractable Fe to dithionite-extractable Fe (active Fe ratio - gives a relative measure of the crystallinity of free iron oxides) is a useful indicator of relative soil development (Blume and Schwertmann, 1969; Alexander, 1974; Alexander and Holawaychuk, 1983; McFadden and Hendricks, 1985). This ratio at first increases with increasing soil development, and then decreases to an approximately constant value. In the Reefton chronosequence a similar trend is shown by the percentage of total Fe and Al that is oxalate-extractable (Campbell, 1975).

Because of the proven usefulness of oxalate-extractable Fe and Al as indicators of relative soil development, emphasis has been placed on trends obtained using this reagent. A few samples were extracted with pyrophosphate and dithionite to indicate the values obtained using these extractants. Ratios of oxalate-extractable Fe and Al to total Fe and Al (determined by XRF) were also calculated. The following abbreviations are used to refer to the various forms of Fe and Al (and Si):

oxalate-extractable	Fe_0, Al_0, Si_0
pyrophosphate-extractable	Fep, Alp
dithionite-extractable	Fe_d, Al_d
total	Fe_t, Al_t

(a) Oxalate chemistry - soils on gentle slopes (Fig. 58). Trends for Fe_0 and Al_0 were very similar. Surface horizon concentrations increased from F1 (0.44% Fe_0, 0.16% Al_0) to F2 (0.63% Fe_0, 0.28% Al_0). Both Fe_0 and Al_0 decreased with depth in these young soils which had surface pH >5.0. In F3 surface horizon concentration of Fe_0 was greater than F2 (0.98%), while Al_0 was similar (0.27%). Incipient podzolisation was evident in F3 with relatively low concentrations of Fe_0 and Al_0 in the E_{agj} horizon (0.57% Fe_0, 0.14% Al_0), and higher concentrations in the B_w horizon (0.92% Fe_0, 0.18% Al_0). F4 had a typical podzol depth distribution, with relatively low concentrations in the A_h and E_{ag} horizons (0.16 to 0.36% Fe_0, 0.23% Al_0) and highest concentrations in the iron pan (12.5% Fe_0 and .90% Al_0) and B_s horizons. F5 had a very similar depth distribution for Fe_0 and Al_0 but with lower concentrations than F4 in the A and E horizons (0.09% Fe_0, 0.19% Al_0). However, concentrations in the iron pan were also lower (5.15% Fe_0 and .52% Al_0). Differences between F4 and F5 may be due to the transformation of poorly-ordered substances to more crystalline forms, or to their dissolution and translocation further down the profile. Maximum concentration of both Fe_0 and Al_0 in F5 occured at greater depth in the profile than in F4. This sequence of five soils shows an initial increase in oxalate extractable Fe and Al in surface horizons (F1 to F3). This is followed by a decrease in A_h and E_{ag} horizon values and a concurrent increase in B horizon values due to translocation of Fe and Al.

F6 had the lowest concentrations of Fe_0 and Al_0 in the A_h and E_{ag} horizons of any of the soils (<0.06% Fe_0, <0.15% Al_0) but did not have a distinctive subsoil maximum since it lacked an iron pan. The lower concentration of Fe_0 and Al_0 may reflect greater loss (or transformation) of Al_0 and Fe_0, or it may result from a slower rate of pedogenesis in soils derived from bedrock.

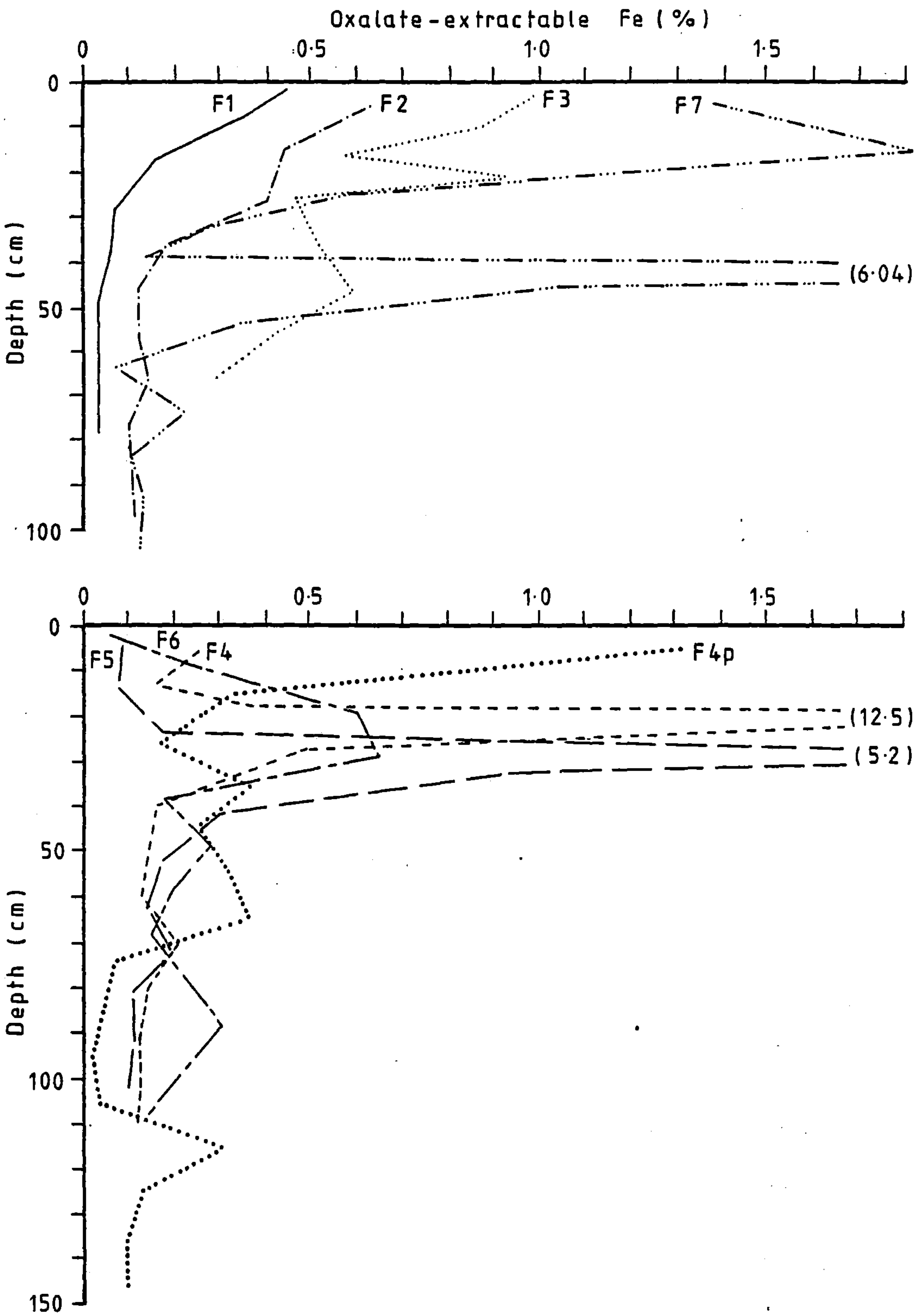


Fig. 58(a): Acid ammonium oxalate-extractable Fe - soils on gentle slopes.

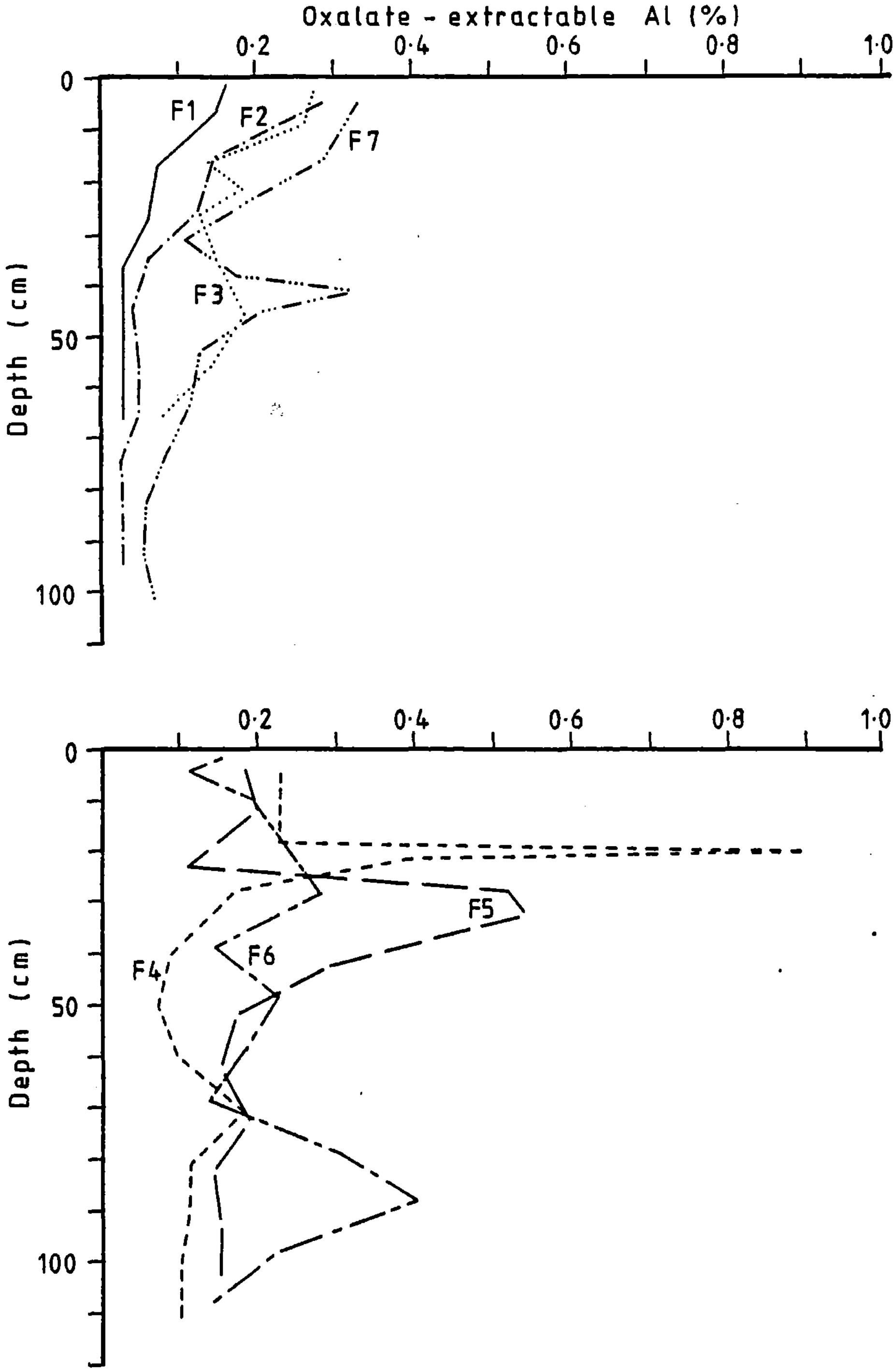


Fig. 58(b): Acid ammonium oxalate-extractable Al - soils on gentle slopes.

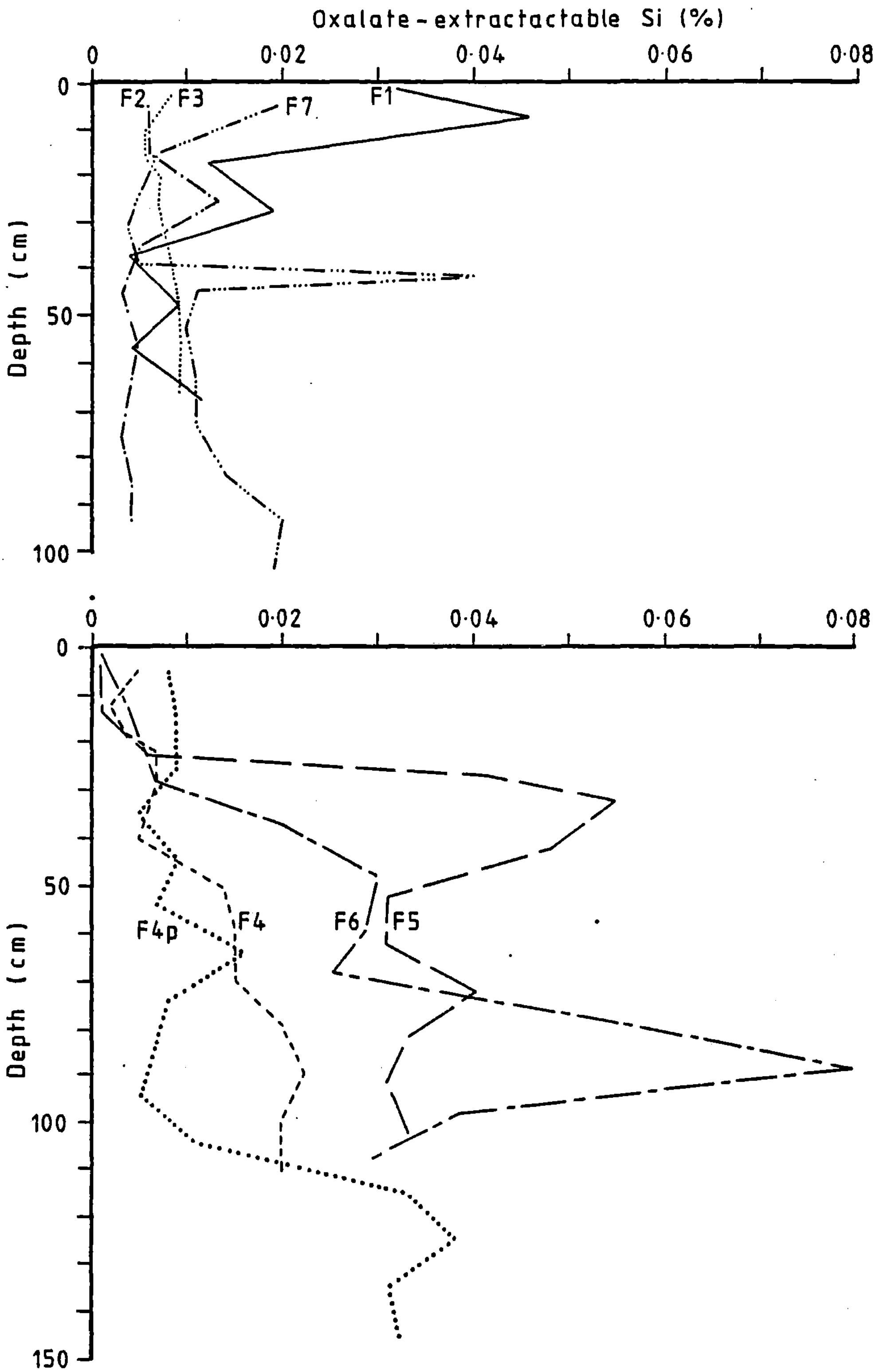


Fig. 58(c): Acid ammonium oxalate-extractable Si - soils on gentle slopes.

F7 (2,500 year old soil from fine textured alluvium) also had a typical podzol depth distribution of Al_0 and Fe_0 with minimum concentrations in the E_{ag} horizon and a maximum for both in the iron pan. This soil was intermediate in oxalate chemistry between F3 and F4 with maximum concentrations in the iron pan, but with high values of Fe_0 and Al_0 in the O_h horizon.

Comparison of F4p and F4 indicated that Al_0 accumulated in poorly drained sites with peat, while Fe_0 did not. Presumably Fe had been mobilised as Fe^{2+} in a reducing environment. Concentration of Al_0 was extremely high (up to 8% on an ignited basis). However this was counteracted by low bulk densities in F4p so that the weight of Al_0 was lower in F4p than in F4. The high concentration of Al_0 probably resulted from the dissolution by ammonium oxalate of Al associated with organic matter (Parfitt, 1980).

For Si_0 the trends with respect to age were not well defined. F1 had a high concentration of Si_0 in the top 30 cm compared to the rest of the soils. In F3, F4 and F6, Si_0 increased down the profile and did not show the subsoil maximum characteristic of the profile depth distribution of Al_0 and Fe_0. B horizon maxima in Si_0 concentration were present in F5 and F7.

b) Oxalate chemistry - soils on steep slopes (Fig. 59). Trends for Al_0 and Fe_0 were again very similar, except that high concentrations of Al_0, but not Fe_0, were present in the peaty surface horizons of S2 and $S5_1$. The young soils (S1 and S2) had low concentrations (<0.90% Fe_0, <0.17% Al_0) and a uniform or slightly decreasing trend with depth. $S3_u$ and $S3_1$ had maximum concentrations of Fe_0 (up to 1.43%) and Al_0 (up to 0.30%) in the A_h or B_w horizons, and decreasing concentrations below the B horizon. However, $S3_1$ also had another maximum in Al_0 concentration in the C horizon where high

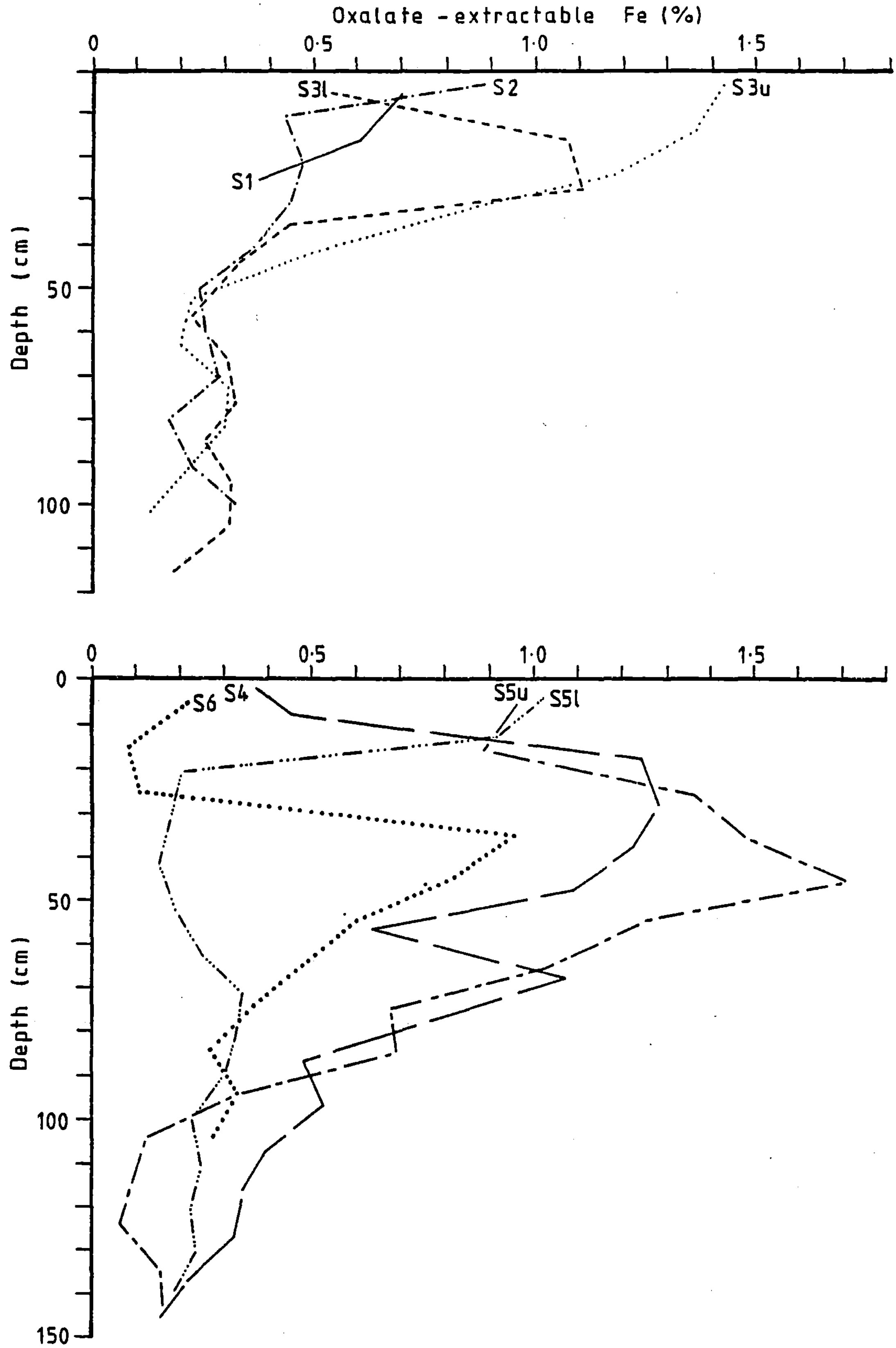


Fig. 59(a) Acid ammonium oxalate extractable Fe - soils on steep slopes.

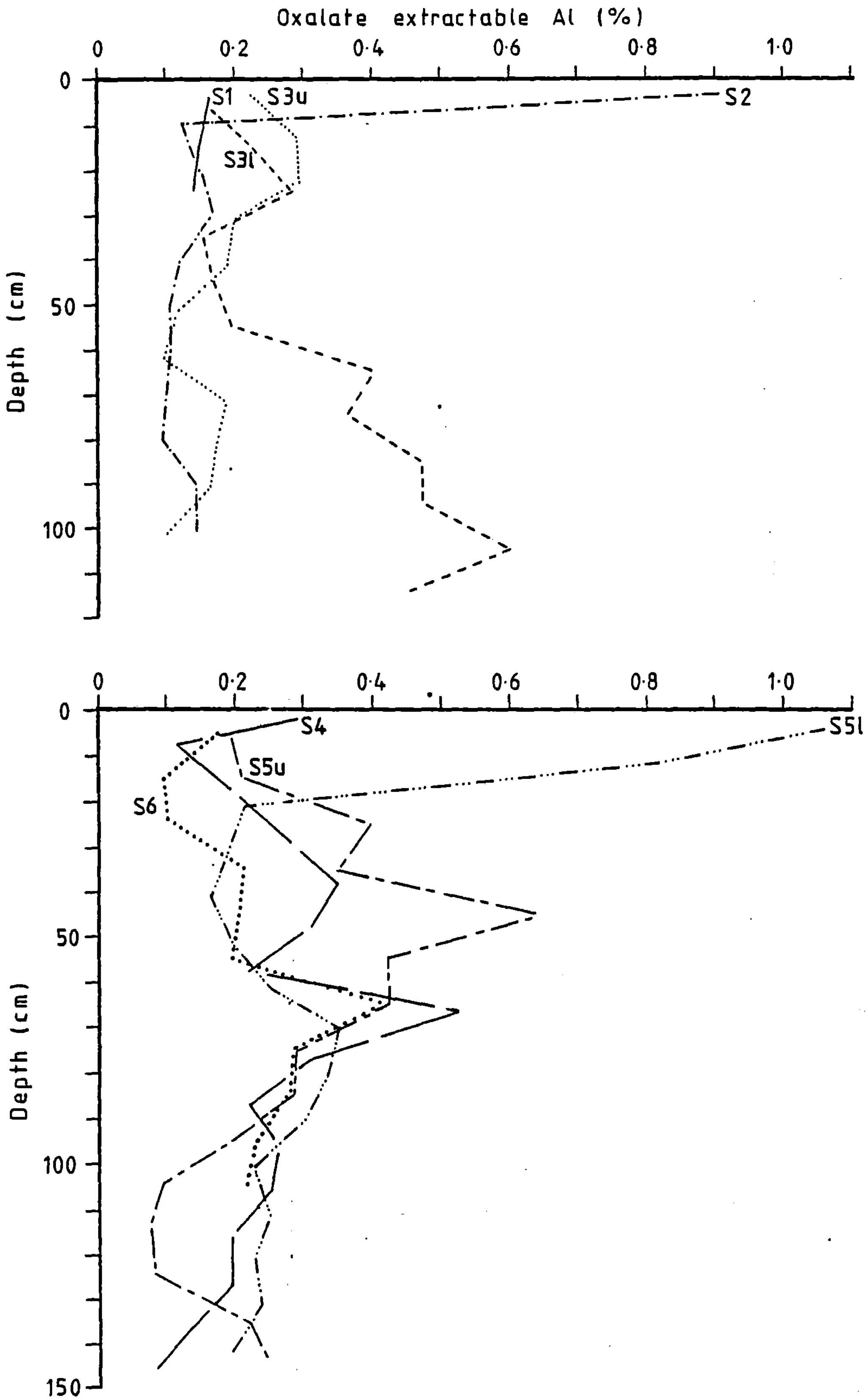
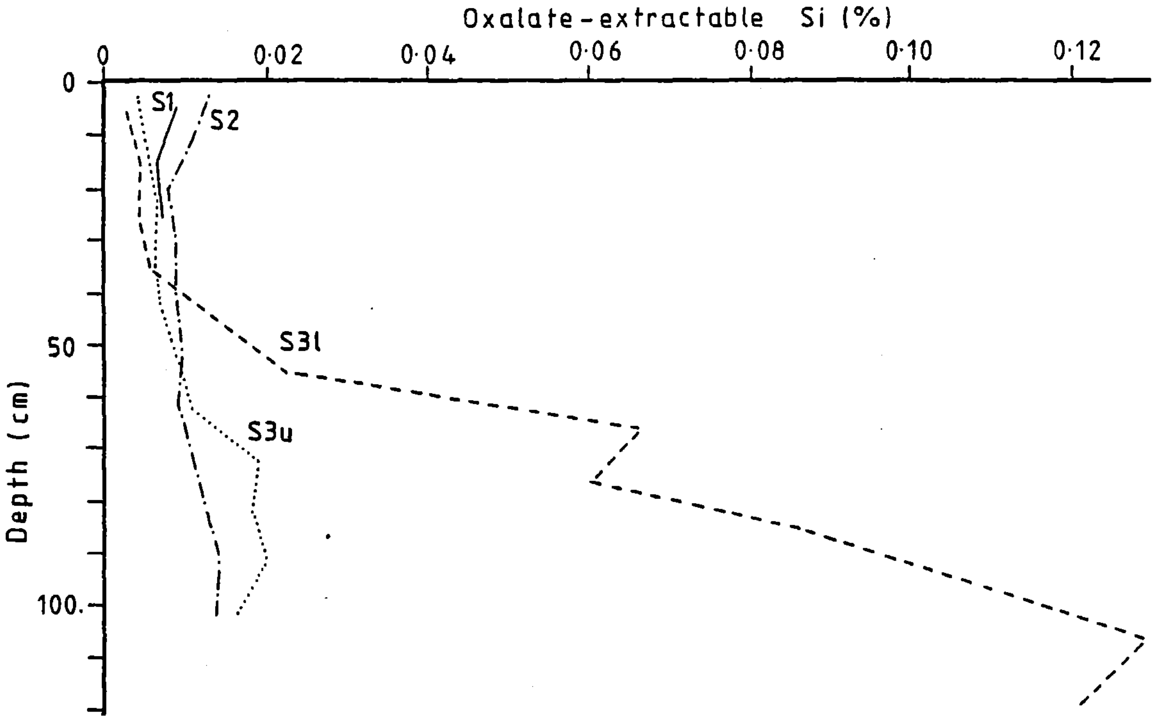


Fig. 59(b): Acid ammonium oxalate extractable Al - soils on steep slopes.



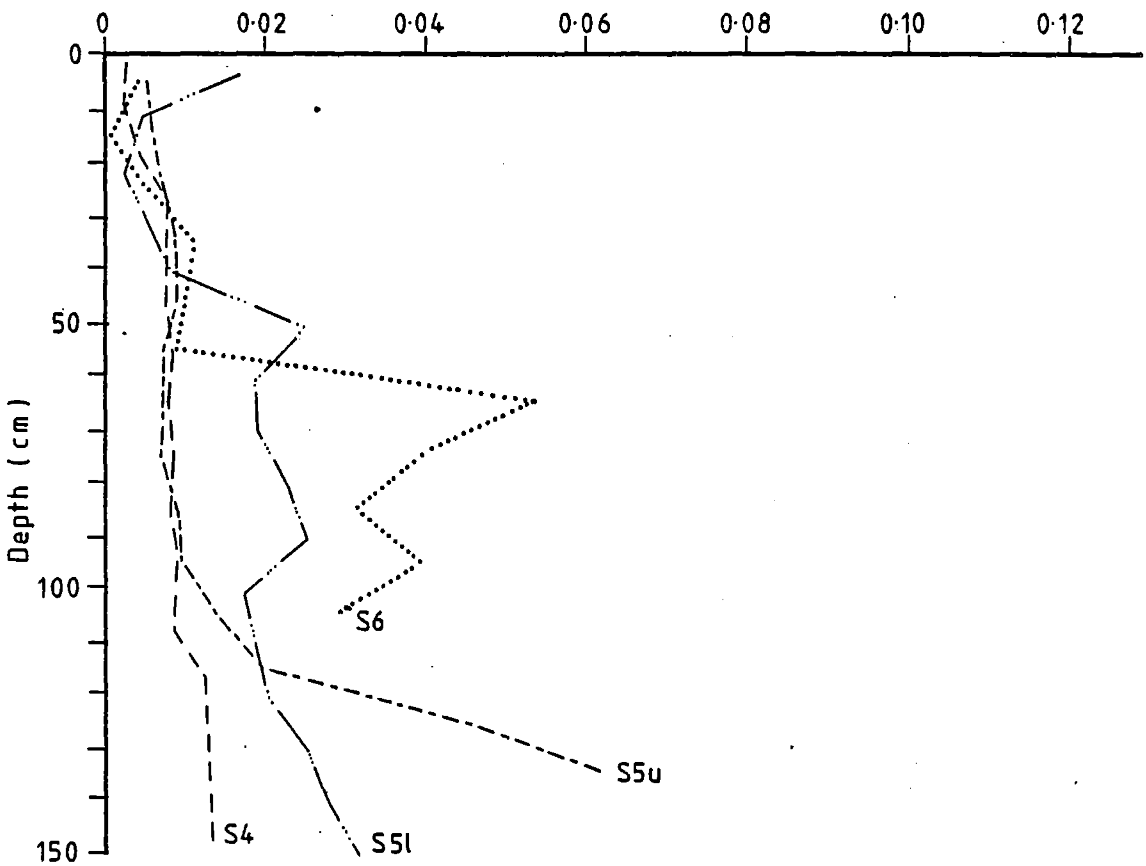


Fig. 59(c): Acid ammonium oxalate extractable Si - soils on steep slopes.

Si_0 concentrations occured. B_w horizon concentrations of Fe_0 were similar in $S3_u$ and $S3_l$, but A_h horizon concentrations were far lower in $S3_l$. This may be indicative of some influence of reducing conditions in the A_h horizon of $S3_l$. S4 was clearly eluviated, with minimum concentrations of Fe_0 (<0.45%) and Al_0 (0.11%) in the A_h and E_{agj} horizons and maximum concentrations in the B_s horizon. Maximum B horizon concentrations of Fe_0 (1.29%) and Al_0 (0.52%) in S4 were similar to those in $S3_u$ and $S3_l$ but high concentrations occurred to greater depth in S4. $S5_u$ had the highest B_s horizon concentrations of Fe_0 (1.73%) and Al_0 (0.65%) of all the soils on steep slopes, but its upper horizons were not as eluviated as S4. $S5_l$ had lower concentrations of Fe_0 and Al_0 throughout the gleyed B_g horizon. S6 had a typical podzol depth distribution of Fe_0 and Al_0. It had the lowest concentrations of all soils of Fe_0 (<0.11%) and Al_0 (0.10%) in the A_h and E_{ag} horizons. Concentrations increased with depth to maximum values in the B_s horizon. The maximum concentration of Fe_0 (0.97%) and Al_0 (0.42%) in S6 was lower than in S4 and $S5_u$.

In general terms there is a progressive change with increasing soil development from low and uniform (or slightly decreasing) concentrations of Fe_0 and Al_0 with depth in recent soils (S1, S2), to maximum concentrations in A_h and B_w horizons in yellow-brown earths (S3), to minimum concentrations in A and E horizons and maximum concentrations in B_s horizons in gley podzols (S4, $S5_u$, S6). Except for $S5_u$, oxalate-extractable Fe and Al values ranked the soils clearly with respect to age.

Trends in Si_0 analyses were not clearly evident. Si_0 was very low except in the C horizon of $S3_l$, which also had anomalously high concentrations of Al_0. In the younger soils (S1, S2) Si_0 concentration was uniform with depth, while in the older soils it tended to increase with depth to a maximum in the C horizon.

(c) Dithionite- and pyrophosphate-extractable Fe and Al and extractable Fe and Al ratios (Table 10).

Trends in amounts of Fe and Al extractable in dithionite and pyrophosphate were generally similar to those obtained for oxalate extracts (Figs. 60, 61). In the two younger soils (F2, S2) concentration decreased with depth, while the two older soils (F4, S6) had maxima in the B_s horizon. The three reagents extracted variable amounts of Fe, with dithionite extracting more than oxalate, and oxalate more than pyrophosphate. All three reagents generally extracted similar amounts of Al.

Active iron ratios (Table 10) generally fell within the range of values recorded by Blume and Schwertmann (1969). The peaty H horizon of S2 gave the highest ratio, indicating strong dissolution of organic-complexed Fe by oxalate. The two young soils (F2, S2) have relatively high values of the activity ratio (>0.5) which decreased with depth. In the two old soils (F4, S6) lower ratios occurred in the A_h and E_{ag} horizons (0.3 to 0.5) and higher ratios occurred in the B_s horizon (0.6 to 0.7). There appeared to be a decrease in the active Fe ratio in surface horizons with increasing age.

The ratios of Fe_0/Fe_t and Al_0/Al_t for the two soil sequences are shown in Figs. 62 to 65. For soils on gently sloping sites Fe_0/Fe_t ratios in surface horizons increased from F1 (0.09) to F3 (0.24). In F4 and F5 ratios in the A_h and E_{ag} horizons decreased (0.06 to 0.12) while B_s horizon ratios increased (0.35 to 0.61). F6 had low values of Fe_0/Fe_t throughout. F7 had a high ratio of Fe_0/Fe_t in the 0 and A_h horizons (0.23 to 0.45), although it had a very low ratio in the E_{ag} horizon (0.04). The ratio of Al_0/Al_t showed similar trends, except that A and E horizon ratios remained reasonably constant from F3 to F6.

For the soils on steep slopes the trends with age were not so clear. S1, S2, $S3_{u,l}$, $S5_{u,l}$ had maximum values of

Sample Code	Fe_p %	Fe_0 %	Fe_d %	Fe_0/Fe_d	Al_p %	Al_0 %	Al_d %
F2.1	0.49	0.63	1.05	0.60	0.26	0.29	0.30
.3	0.21	0.41	0.70	0.58	0.12	0.14	0.16
.10	0.03	0.11	0.22	0.50	0.02	0.03	0.04
F4.1	0.21	0.26	0.45	0.56	0.16	0.23	0.14
.2	0.10	0.16	0.39	0.42	0.15	0.23	0.12
.3	0.21	0.36	1.09	0.33	0.17	0.23	0.18
.5	1.61	1.82	2.67	0.68	0.41	0.39	0.47
.12	0.03	0.13	0.28	0.46	0.06	0.11	0.05
S2.1	0.67	0.89	1.09	0.82	1.05	0.91	1.00
.2	0.33	0.45	0.70	0.64	0.15	0.11	0.13
.4	0.39	0.45	0.74	0.61	0.17	0.17	0.17
.7	0.21	0.26	0.54	0.47	0.14	0.11	0.12
.11	0.23	0.34	0.52	0.65	0.16	0.14	0.15
S6.2	0.14	0.08	0.23	0.33	0.14	0.09	0.15
.3	0.13	0.11	0.28	0.41	0.08	0.10	0.05
.4	0.92	0.97	1.45	0.67	0.22	0.21	0.26
.5	0.94	0.83	1.36	0.61	0.25	0.21	0.35
.7	0.42	0.47	0.82	0.57	0.39	0.41	0.34
.11	0.22	0.29	0.55	0.52	0.20	0.21	0.20

TABLE 10: Comparison of pyrophosphate-, oxalate-, and dithionite-extractable iron and aluminium for selected samples of F2, F4, S2 and S6.

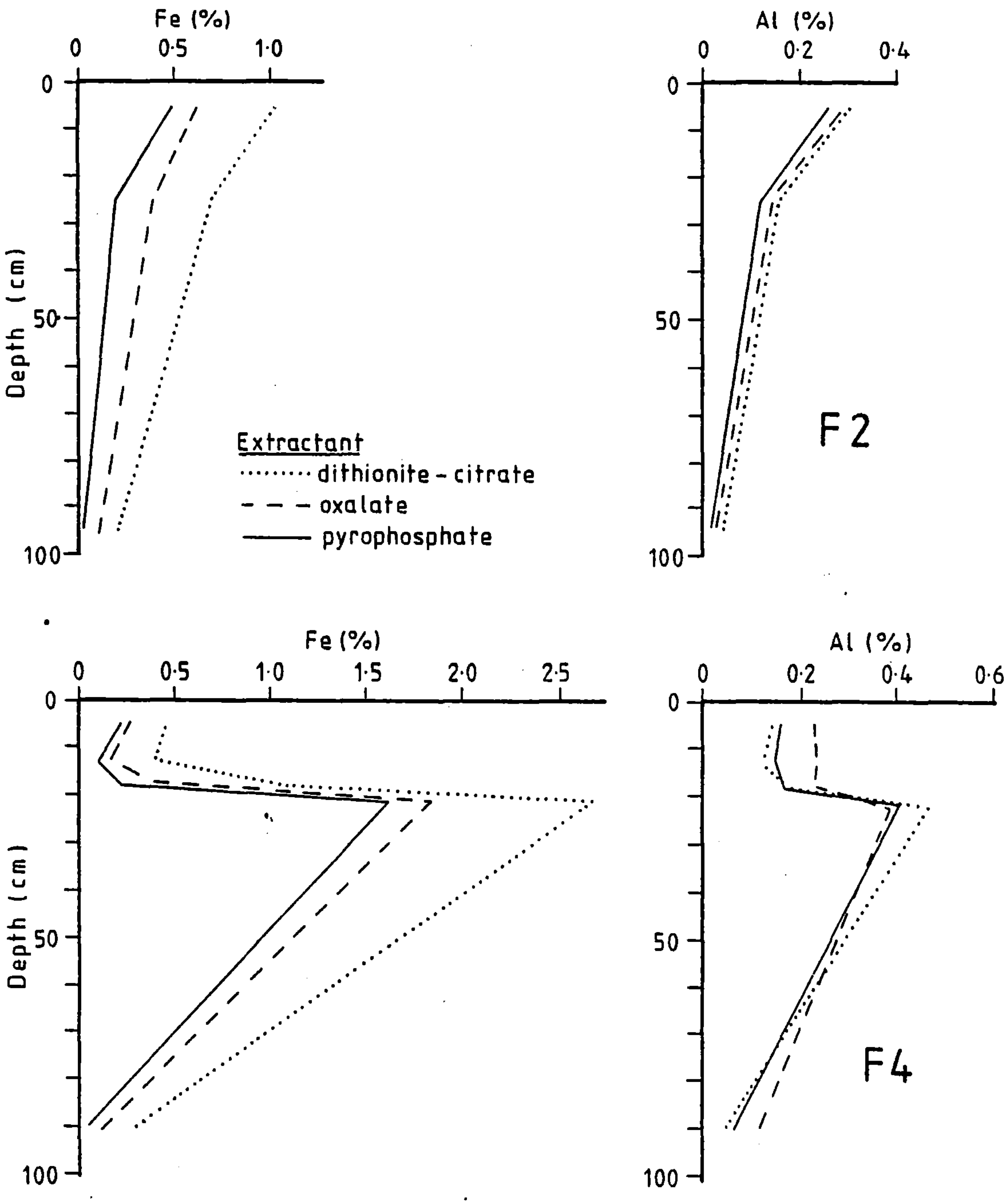


Fig. 60: Pyrophosphate-, oxalate-, and dithionite-citrate extractable Fe and Al in selected samples of F2 and F4.

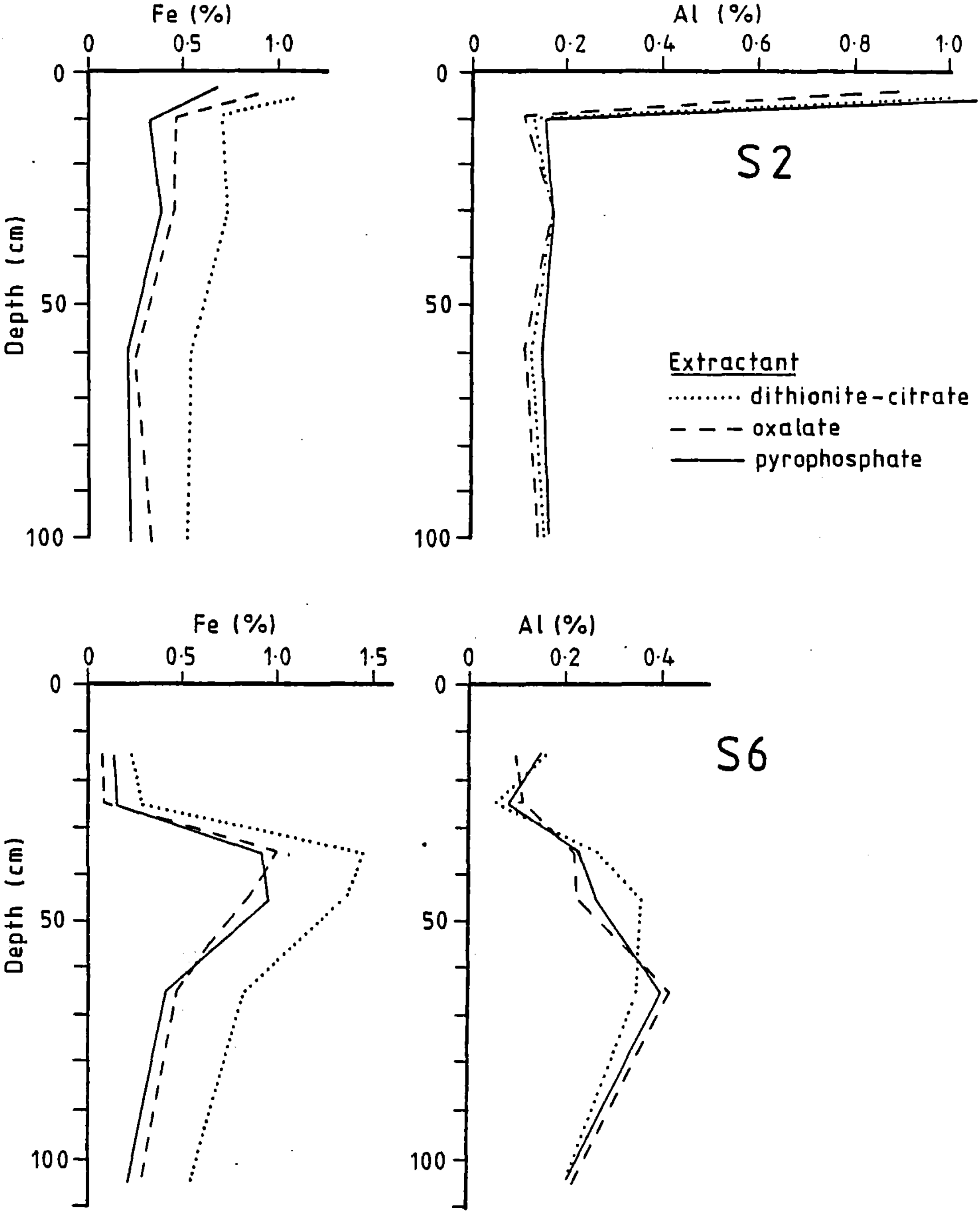


Fig. 61: Pyrophosphate-, oxalate-, and dithionite-citrate extractable Fe and Al in selected samples of S2 and S6.

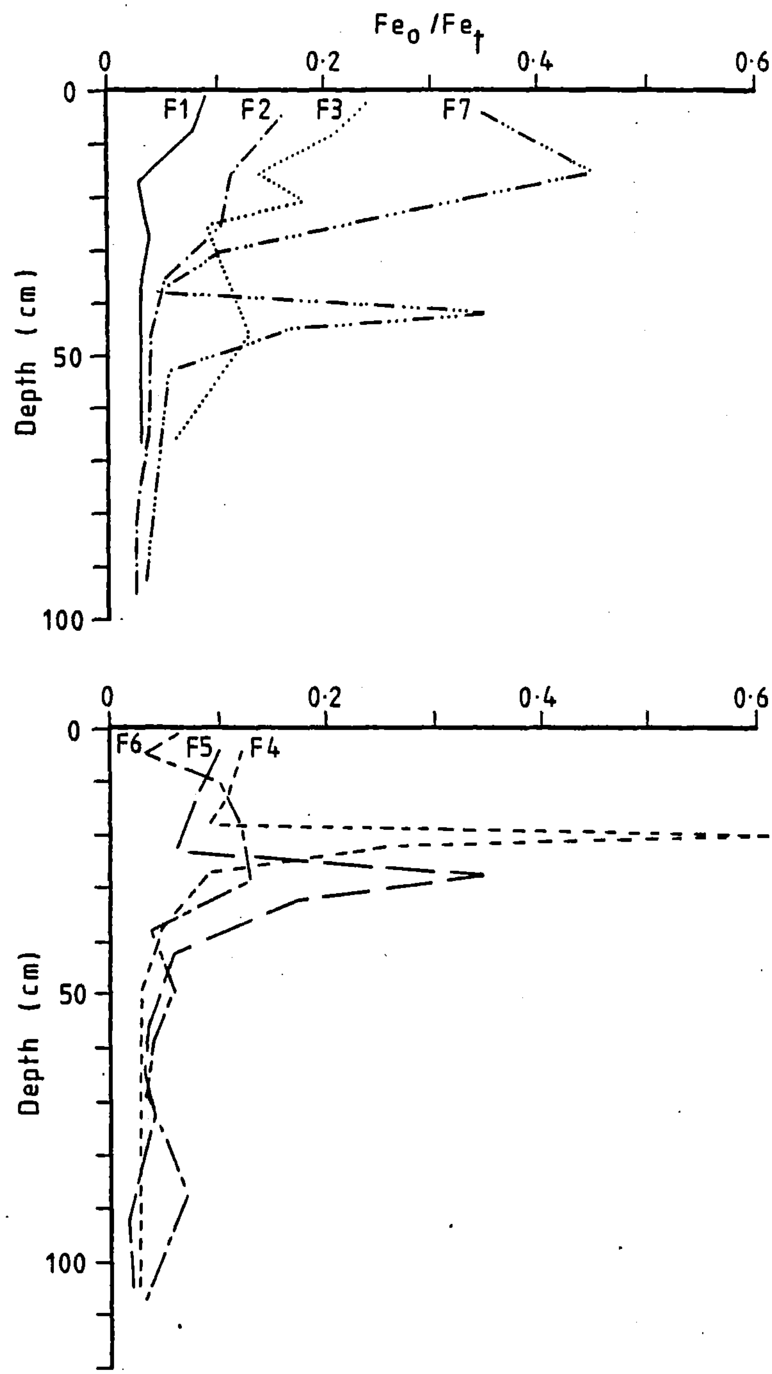


Fig. 62: Ratio of oxalate-extractable Fe to total Fe - soils on gentle slopes.

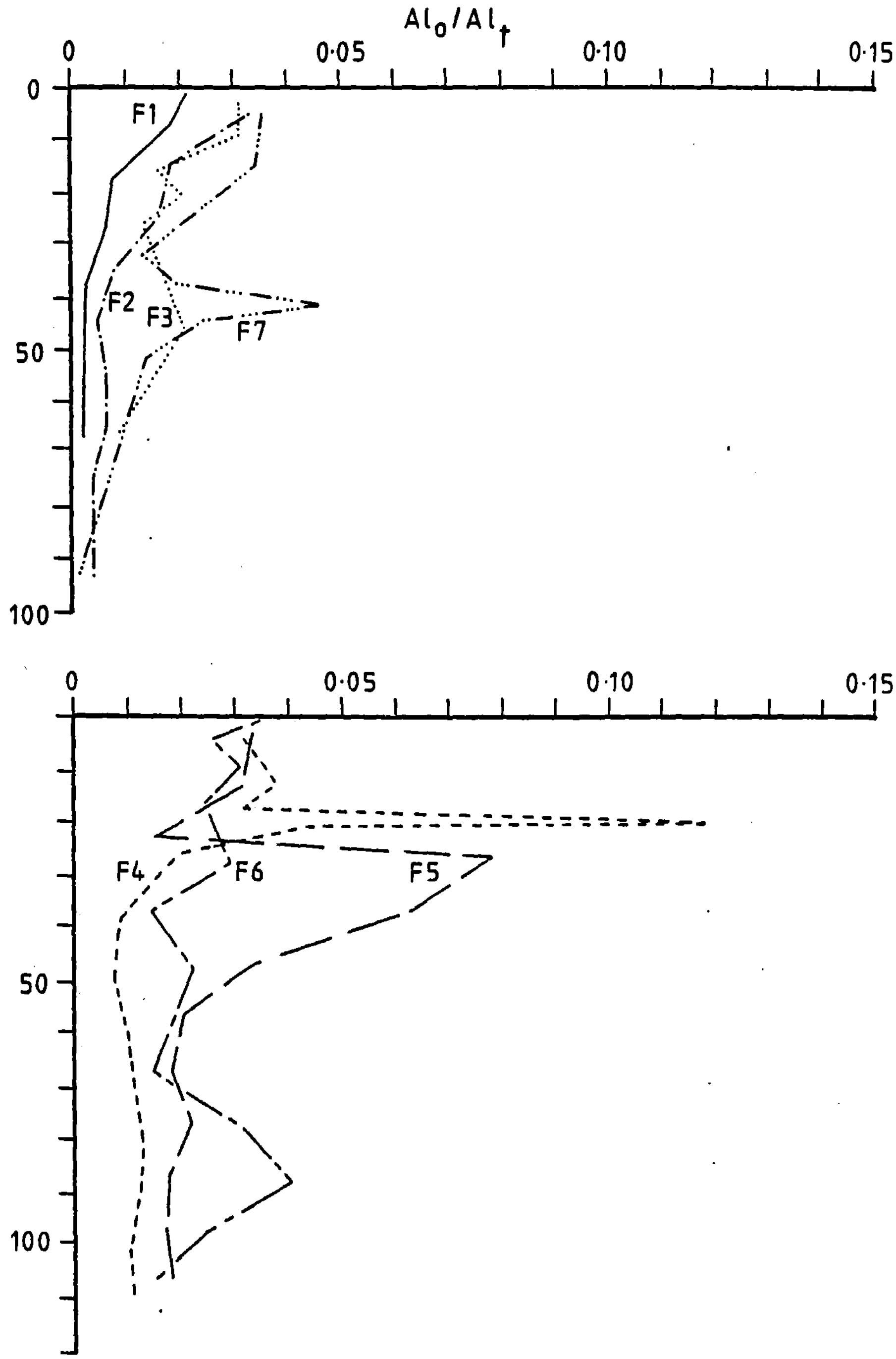


Fig. 63: Ratio of oxalate-extractable Al to total Al - soils on gentle slopes.

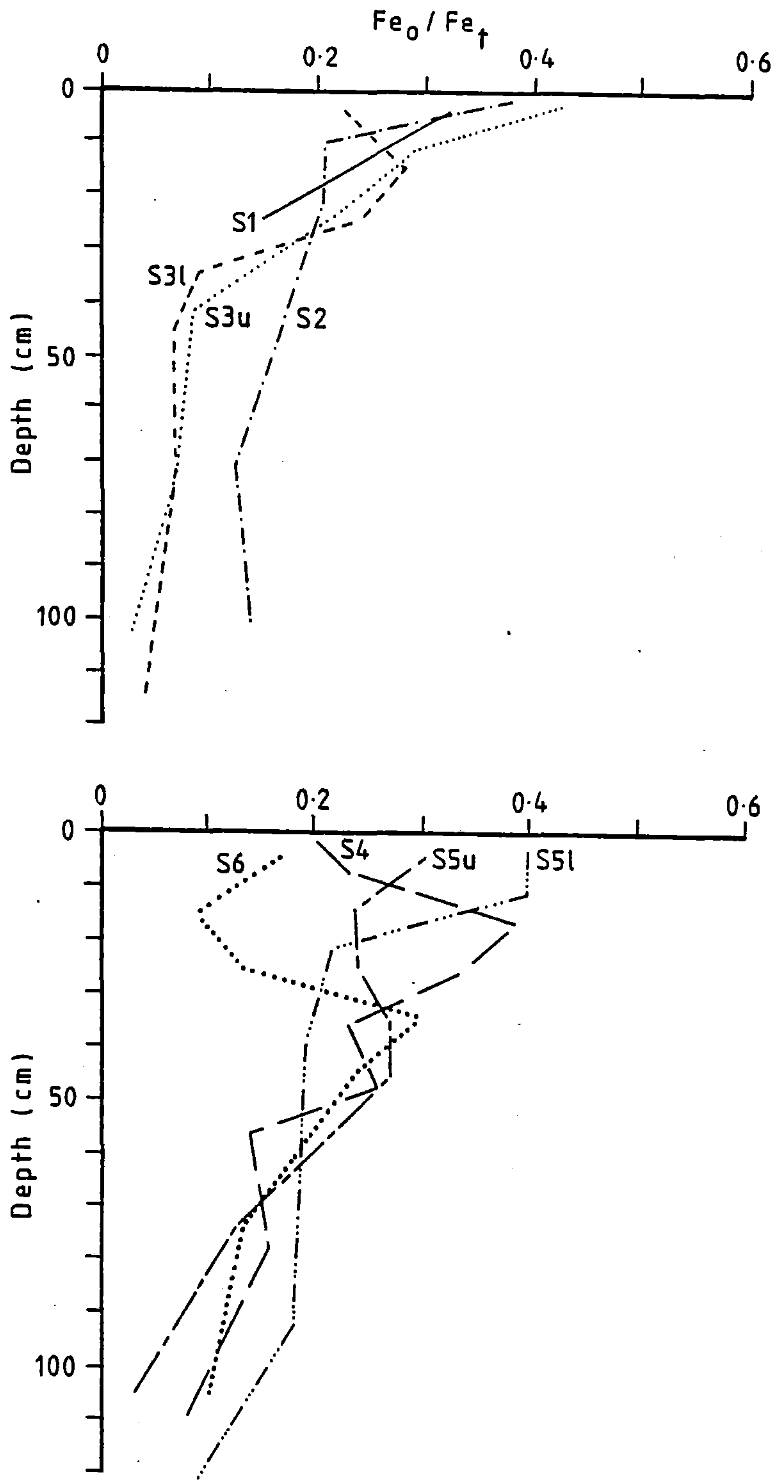


Fig. 64: Ratio of oxalate-extractable Fe to total Fe - soils on steep slopes.

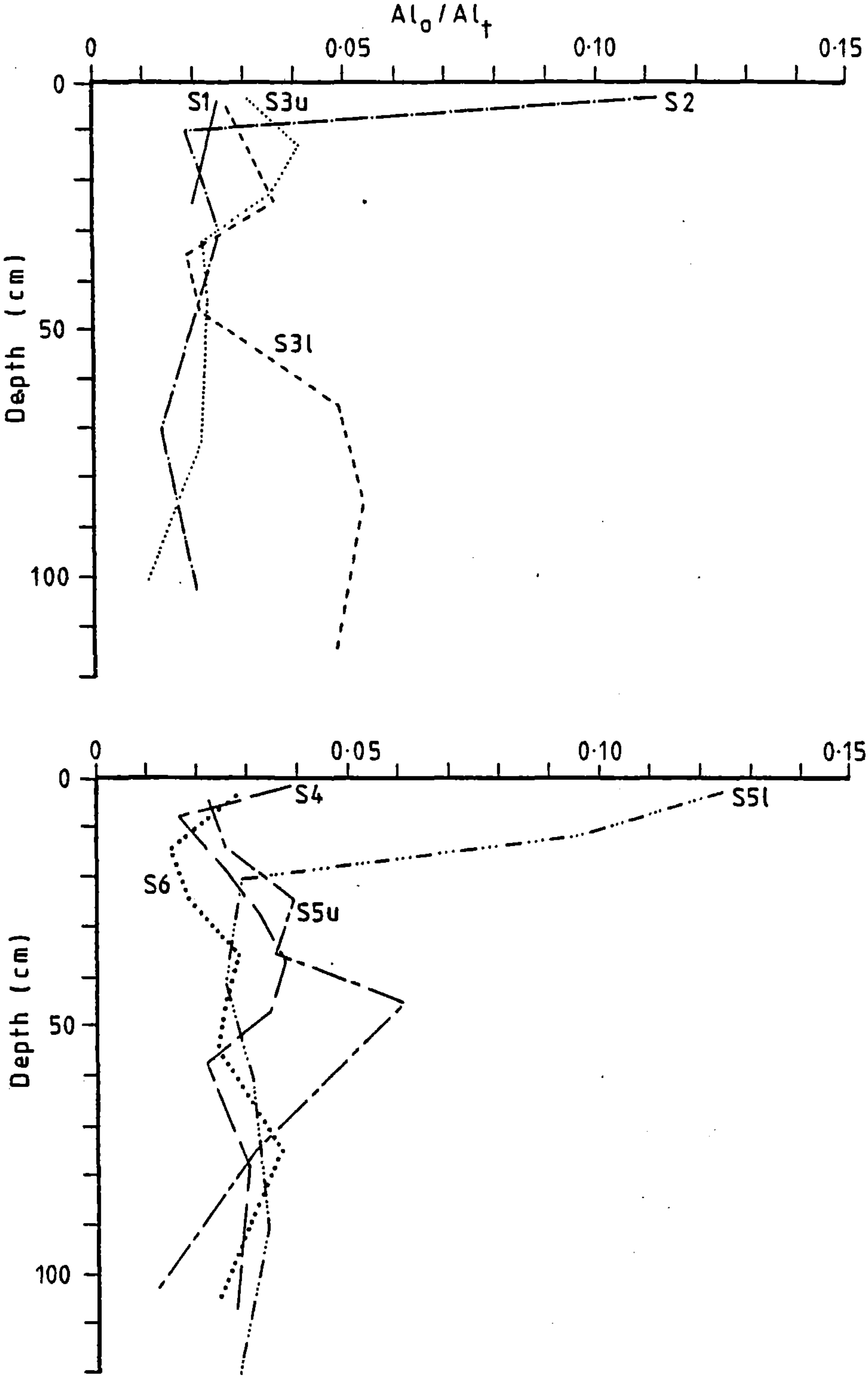


Fig. 65: Ratio of oxalate-extractable Al to total Al - soils on steep slopes.

Fe_0/Fe_t in the A_h and B_w horizons, and values decreased with depth. S4 and S6 both had low values in the A_h and E_{ag} horizons and maximum values in the B_s horizon. Ratios of Al_0/Al_t were influenced by organic matter, with highest values in the H horizons of S2 and $S5_1$. This was a result of the strong dissolution by ammonium oxalate of Al associated with organic matter. No clear trends with respect to age were obvious in the Al_0/Al_t ratio.

6.6.5 Total element analysis

Total element analysis has been widely used to characterise soil profiles and parent materials, and to quantitatively evaluate soil development (see for examples Rode, 1970; Evans and Adams, 1975a, b; Campbell, 1975; Evans, 1978; Muir and Logan, 1982; Chittleborough *et al*, 1984). The quantitative evaluation of profile development involves determination of the properties and uniformity of the parent material, and evaluation of the change in mobile constitutents during soil development (with reference to an inert internal standard).

To quantitatively evaluate changes in mobile constituents during soil development, three assumptions must be made. These are:

(a) the material from which the soil has developed was originally uniform in composition. Variability in chemical composition of the parent material should not approach that produced by pedogenesis;

(b) the C horizon is nearest in composition to that of the original material;

(c) the substance chosen as an internal standard is immobile and inert.

These assumptions are only partially testable.

Uniformity of parent material can be evaluated using stratigraphical and analytical techniques. The presence of more than one parent material in a soil profile can be established from stratigraphy (Butler, 1959), morphology, or from the depth distribution of soil properties, which commonly exhibit simple trends in uniform parent materials (Brewer, 1964; Wang and Arnold, 1973). A variety of analytical methods have been used to test for uniformity of parent material including total mineralogical analysis, particle size distribution of resistant minerals of the non-clay fraction, particle size distribution of the whole non-clay fraction, and chemical composition of the non-clay fraction (Marshall and Haseman, 1942; Barshad, 1964; Brewer, 1964; Chapman and Horn, 1968; Sudom and St Arnaud, 1971; Wang and Arnold 1973; Evans and Adams, 1975a; Evans, 1978; Smeck and Wilding, 1980; Thompson *et al*, 1981). All involve consideration of the more resistant components of the soil, that have been relatively unaffected by soil processes.

In this study, uniformity of parent material within a soil has been examined from considerations of stratigraphy, soil morphology and the depth distribution of soil chemical properties, particularly Zr concentration. A number of the soils exhibit textural breaks, but chemical composition of both fine textured and coarse textured alluvium/colluvium is similar, since both are derived from quartzofeldspathic bedrock.

Quantitative determination of the extent and manner in which soil parent material is altered is made by assuming that certain minerals (referred to as internal standards or index minerals) remain unaltered and immobile during soil development. Assessments of change can then be made relative to the internal standard. While no mineral is completely stable under all conditions, the aim is to choose the most stable and immobile constituent that yields the least possible error in evaluating profile development. Much attention has been given to heavy minerals (zircon, tourmaline, rutile and garnet), quartz, albite, orthoclase, and tri-acid resistant residue (Barshad, 1964; Brewer, 1964; Sudom and St Arnaud, 1971; Evans and Adams, 1975a; Evans, 1978; Chittleborough *et al*, 1984).

Zircon is generally regarded as one of the most resistant heavy minerals. It has been widely used as an internal standard (Marshall and Haseman, 1942; Khan, 1959; Barshad, 1964; Green, 1966; Sudom and St Arnaud, 1971; Wang and Arnold, 1973; Evans and Adams, 1975a; Smeck and Wilding, 1980; Thompson *et al*, 1981) although it has been shown to weather appreciably in a number of environments (Carroll, 1953; Raeside, 1959; Wild, 1961). Commonly it is assumed that zirconium occurs exclusively in zircon, and hence analysis of total zirconium provides a surrogate measure of the mineral zircon. In the Cropp study, zirconium concentration (of the <2 mm fraction) was used as an internal standard since zirconium was the least mobile of the elements analysed, and was easily determined by XRF. Zircon grains, were not examined for signs of weathering.

There are several ways to calculate changes due to pedogenesis. These depend on the analytical data available, and whether absolute or relative changes are required (Barshad, 1964; Brewer, 1964; Rode, 1970; Wang and Arnold, 1973; Evans and Adams, 1975a, b; Walker and Green, 1976; Evans, 1978; Smeck and Wilding, 1980; Muir and Logan, 1982). Ideally changes in soil constituents should be compared on the basis of weight per unit volume, because of volume changes during pedogenesis (Barshad, 1964; Campbell, 1975; Evans and Adams, 1975a; Evans, 1978). Such comparisons were not practical in this study. Only comparative losses and gains have been evaluated.

Evaluation of profile development within soils, and comparisons between soils, was made by calculating eluvial-illuvial (E-I) coefficients (Rode, 1970; Muir and Logan, 1982). This method is relatively simple requiring major element composition, organic content, and internal standard content to calculate relative gains and losses of major elements. These coefficients are calculated from the formula:

$$\text{E-I coefficient (\%)} = \left[\left(S_h * \frac{X_{PM}}{X_h} \middle/ S_{PM}\right) - 1\right] * 100$$

where S_h = concentration of constituent S in horizon h

S_{PM} = concentration of constituent S in parent material

X_{PM} = concentration of internal standard in parent material

X_h = concentration of internal standard in horizon h

The coefficients are converted into percentage coefficients, with a negative coefficient indicating relative loss and a positive coefficient indicating relative gain. They measure the intensity of eluviation/illuviation processes within the soil, and provide a means of assessing relative degrees of profile development between soils with differing original chemical composition. Coefficients were calculated for Ca, P, Fe, Al and Si.

Rearrangement of the above formula indicates that it is a method for comparing total element ratios between the parent material (C horizon) and upper horizons. For example, consider the coefficient calculated for Si, with Zr as the internal standard:

$$\text{E-I coefficient} = \left(Si_h * \frac{Zr_{PM}}{Zr_h} \middle/ Si_{PM} \right) - 1$$

$$= \left(\frac{Si_h}{Zr_h} \middle/ \frac{Si_{PM}}{Zr_{PM}} \right) - 1$$

The use of total element ratios offsets to some extent the impact of volume changes during pedogenesis as both mobile constituents and internal standard are similarly affected by bulk density changes.

a) Soils on gentle slopes.

Total Ca (Fig. 66a) in surface horizons declined progressively from F1 (1.02%) to F5 (0.32%). F6 (old soil from bedrock) had a similar soil profile depth trend to F4 and F5, supporting the interpretation of the age of this soil. F7 had a high concentration of total Ca throughout the profile, probably as a result of an initially high Ca content. C horizon concentrations in the oldest soils (F4, F5, F6) were lower than in F1 to F3, indicating

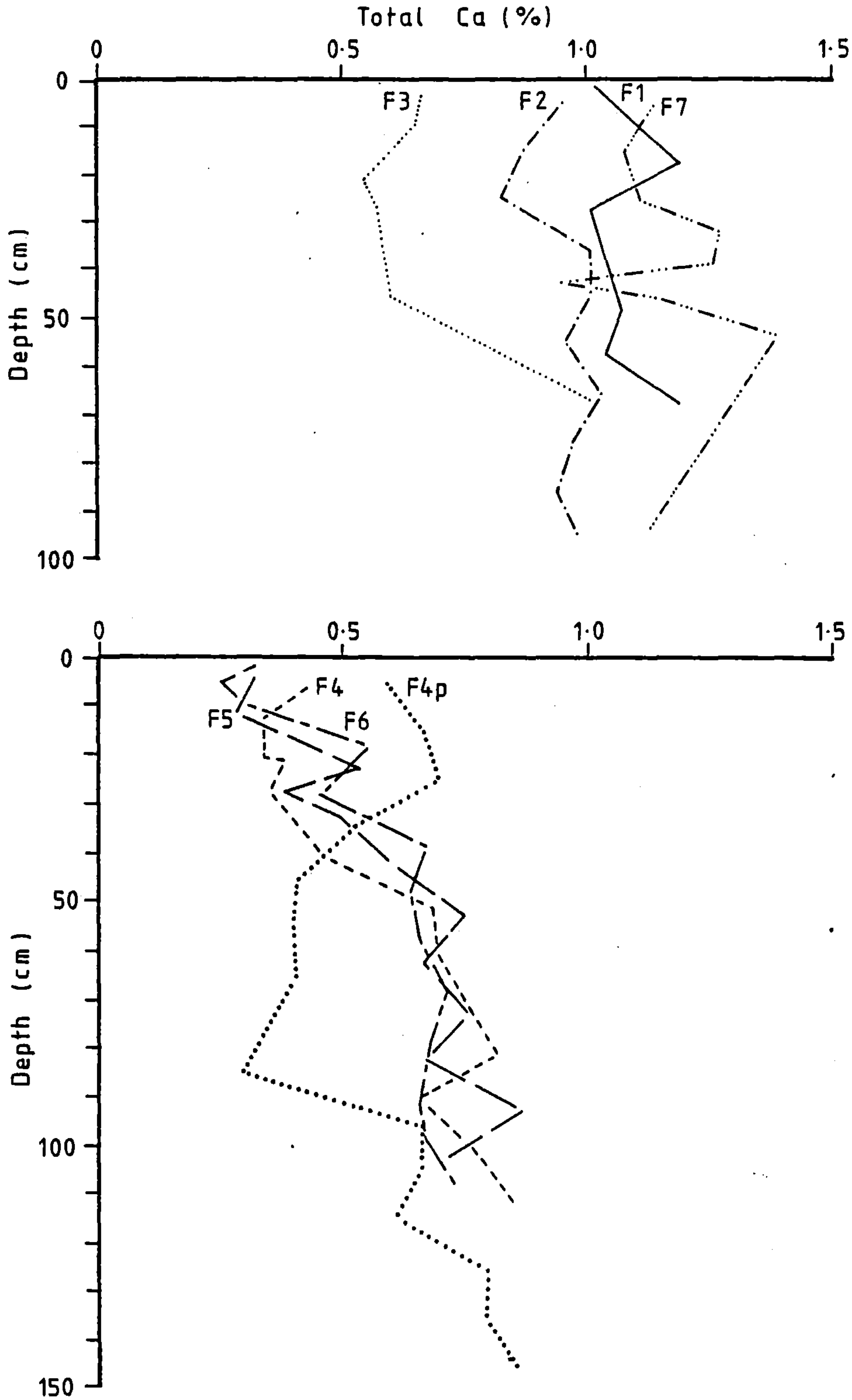


Fig. 66(a): Total Ca - soils on gentle slopes.

either there were different Ca concentrations in the parent materials, or that the depth of weathering was beyond the sampling depth. Each soil showed little variation in C horizon values of total Ca.

Total P analyses (Fig. 66b) showed high concentrations in the A horizons, associated with organic matter, lower concentrations in the E and B horizons and increased concentrations in the C horizons. There was a general decrease in subsoil total P concentration from F1 (0.09%) to F5 (0.02%). F6 had a similar soil profile depth trend to the older soils (F4 and F5), while F7 had the lowest values in the E_{ag} horizon of any soil. F4p had an extremely high concentration of total P associated with the thick organic profile. C horizon concentrations of total P varied from 0.05 to 0.12%.

Total Fe (Fig. 66c) had a uniform depth trend in the recent soils (F1, F2), and increased slightly with depth in F3. In F4 and F5 there were low concentrations in A_h and E_{ag} horizons and a prominent subsoil maxima in the iron pans. B horizon maximum concentration of total iron was greater for F4 (20.62%) than for F5 (14.84%), however minimum concentrations in the A_h and E_{ag} horizons were lower in F5. While F6 did not have a high concentration of total Fe in the subsoil, it had extremely low concentrations in the A_h and E_{ag} horizons. F7 had a similar depth trend of total Fe to F4 and F5, except for higher values in the surface organic horizons of F7. Total Fe is very low throughout the organic surface horizon of F4p compared to F4, as a result of loss of Fe under reducing conditions. All soils had similar, and uniform, concentrations in their C horizons indicating reasonable parent material uniformity.

Depth trend of total Al (Fig. 66d) was uniform in F1, F2 and F3, whereas F4 and F5 showed marked increases in total Al with depth. There was a slight increase in total Al in

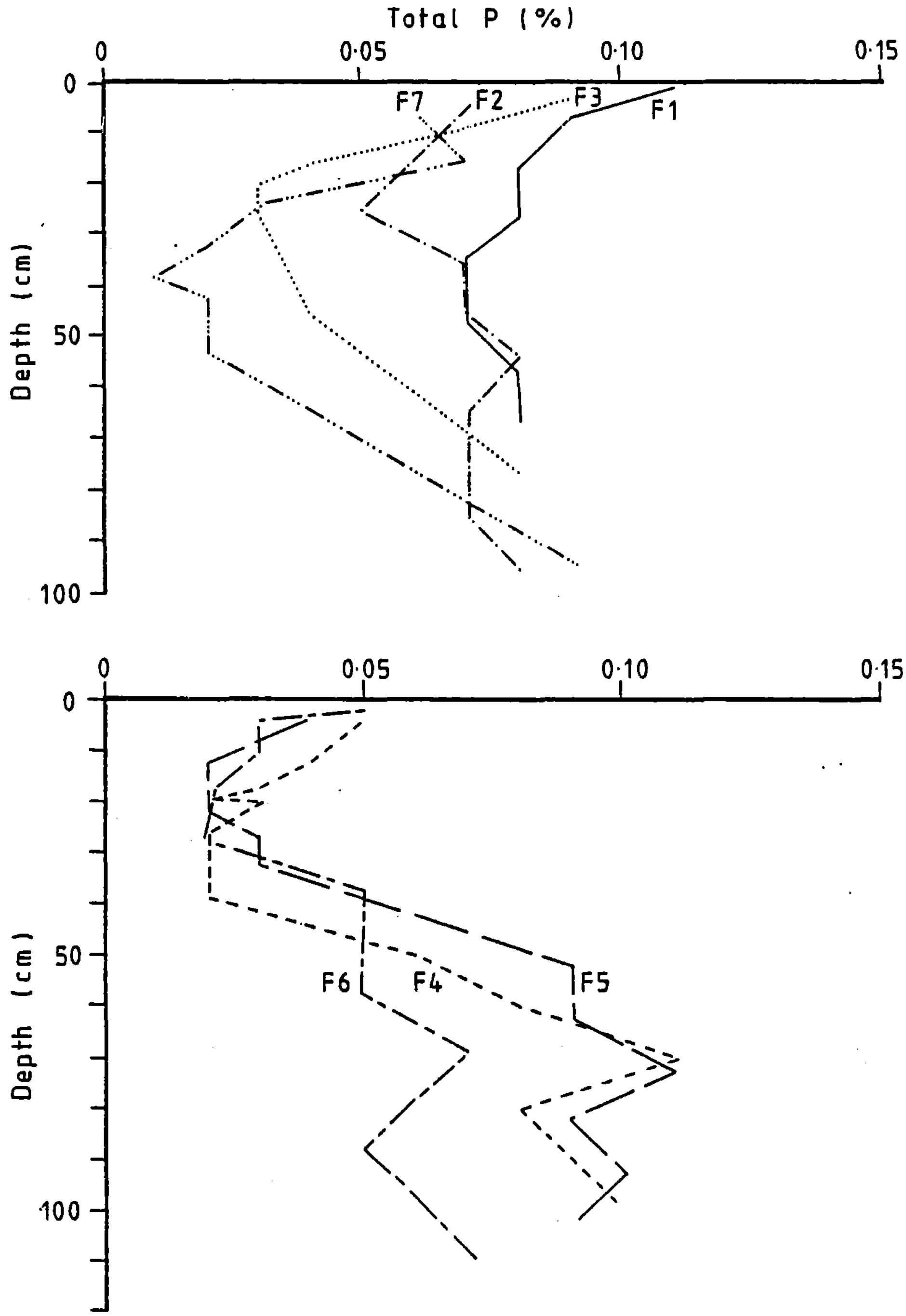


Fig. 66(b): Total P - soils on gentle slopes.

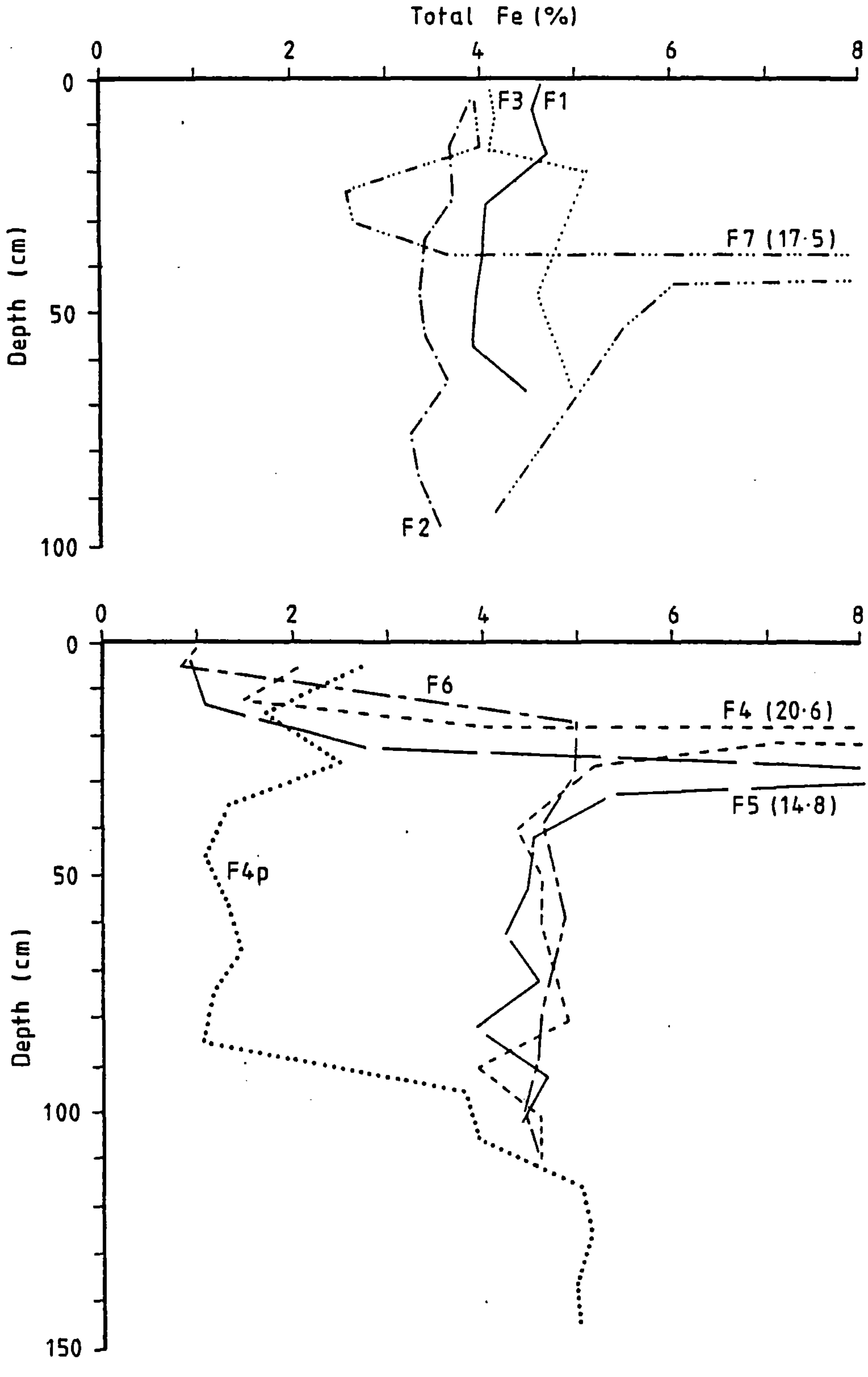


Fig. 66(c): Total Fe - soils on gentle slopes.

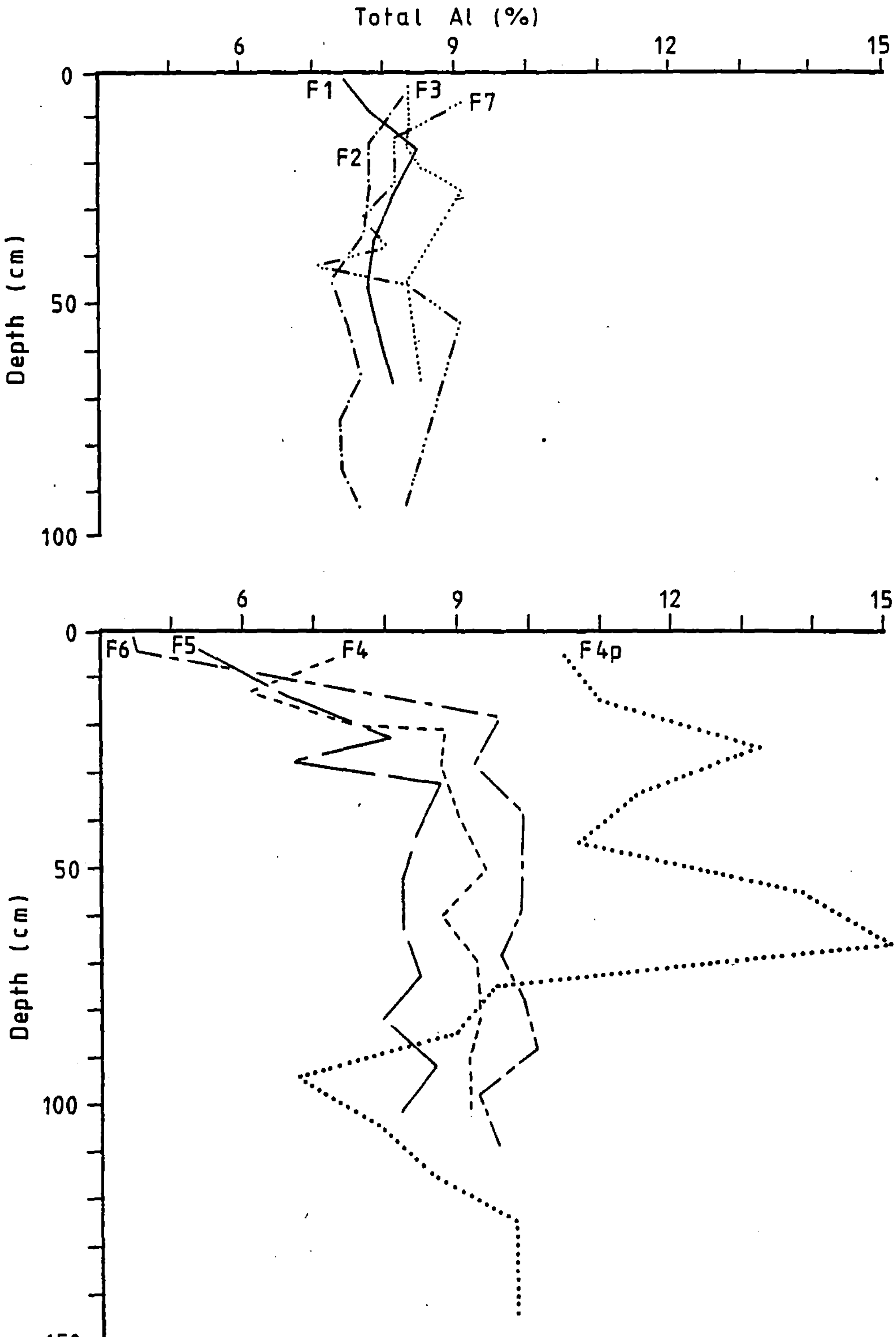


Fig. 66(d): Total Al - soils on gentle slopes.

surface horizons from F1 (7.42%) to F3 (8.35%), and a progressive decrease from F3 to F5 (5.41%). F6 had a similar soil profile depth trend to F4 and F5, with lower surface horizon values of total Al (4.44%). F7 had a similar depth trend to the young soils (F1 to F3). The organic surface horizon of F4p had the highest concentrations of total Al of all soils. There was a considerable range in the C horizon concentrations of different soils (7.5 to 10%) but the C horizon of each soil showed little variation in total Al.

Trends in total Si were opposite to those for Ca, P, Fe, and Al, i.e. Si concentrations tended to increase with age as Si accumulated relative to other elements (Fig. 66e). F1 to F3 had relatively uniform depth trends of total Si (31 to 33%). From F4 to F6 there was a progressive increase in total Si concentration in the A_h and E_{ag} horizons (35% to 40%), while minimum concentrations were recorded in iron pans. F7 had a similar profile depth trend to the older soils (F4 to F6) but surface horizon values were lower (31 to 34%). While total Si contents of the C horizons of the soils varied (30 to 34%), concentrations within any one C horizon were relatively uniform.

Total Ca and P were useful for distinguishing between young soils (F1 to F3), but not for older soils. Total Fe, Al and Si were poor for distinguishing between young soils, but were useful for the older soils (F4 to F6).

Elemental ratios of Si/Al, Si/Fe, and Al/Fe (Fig. 67) showed the younger soils (F1 to F3, F7), with uniform depth distribution, differed distinctly from the older soils (F4 to F6) which had higher values for all three ratios in surface horizons. The C horizon of each soil had uniform values of these ratios and there was little variation between soils in C horizon values. This indicated reasonably uniform parent materials, both within

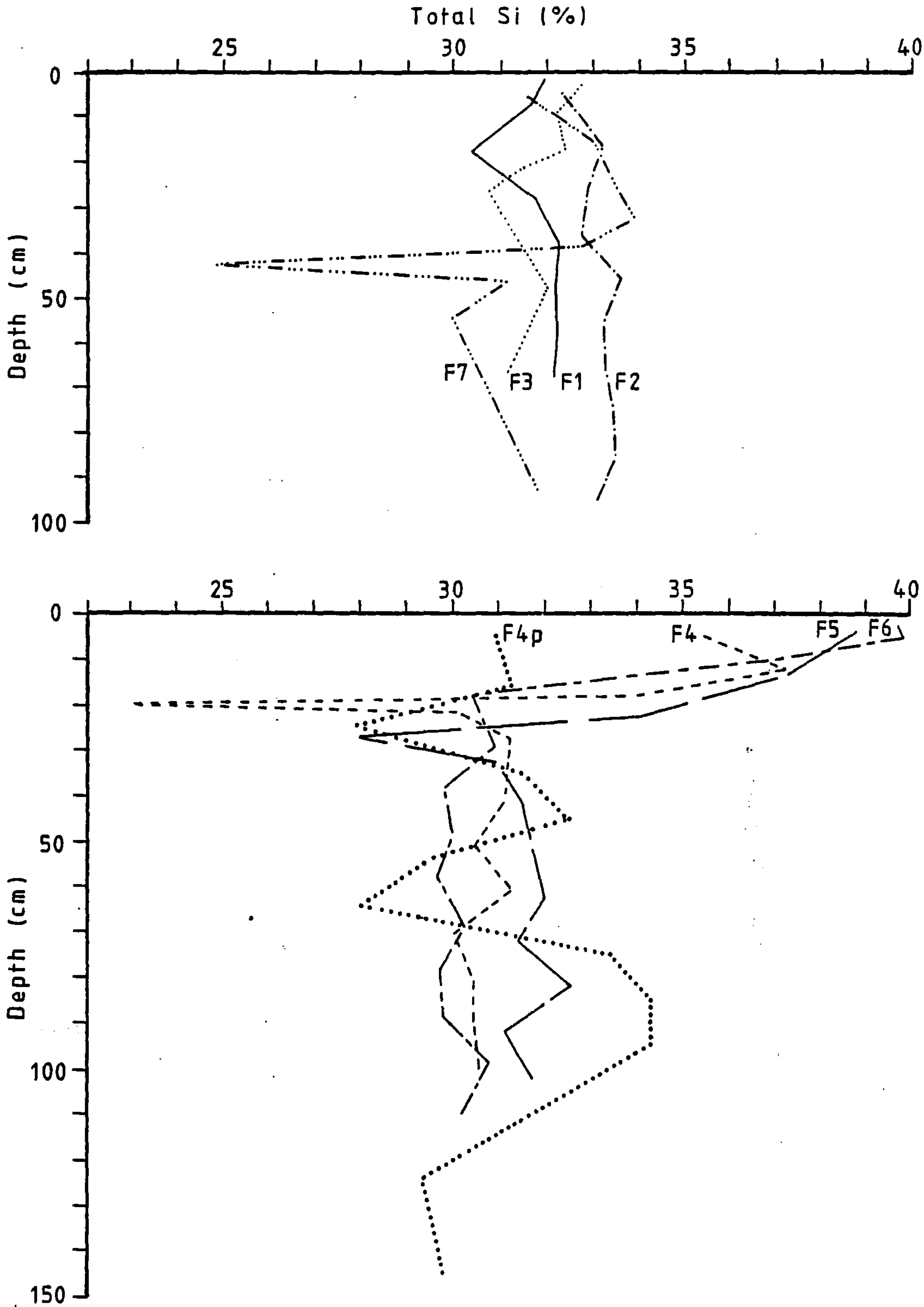


Fig. 66(e): Total Si - soils on gentle slopes.

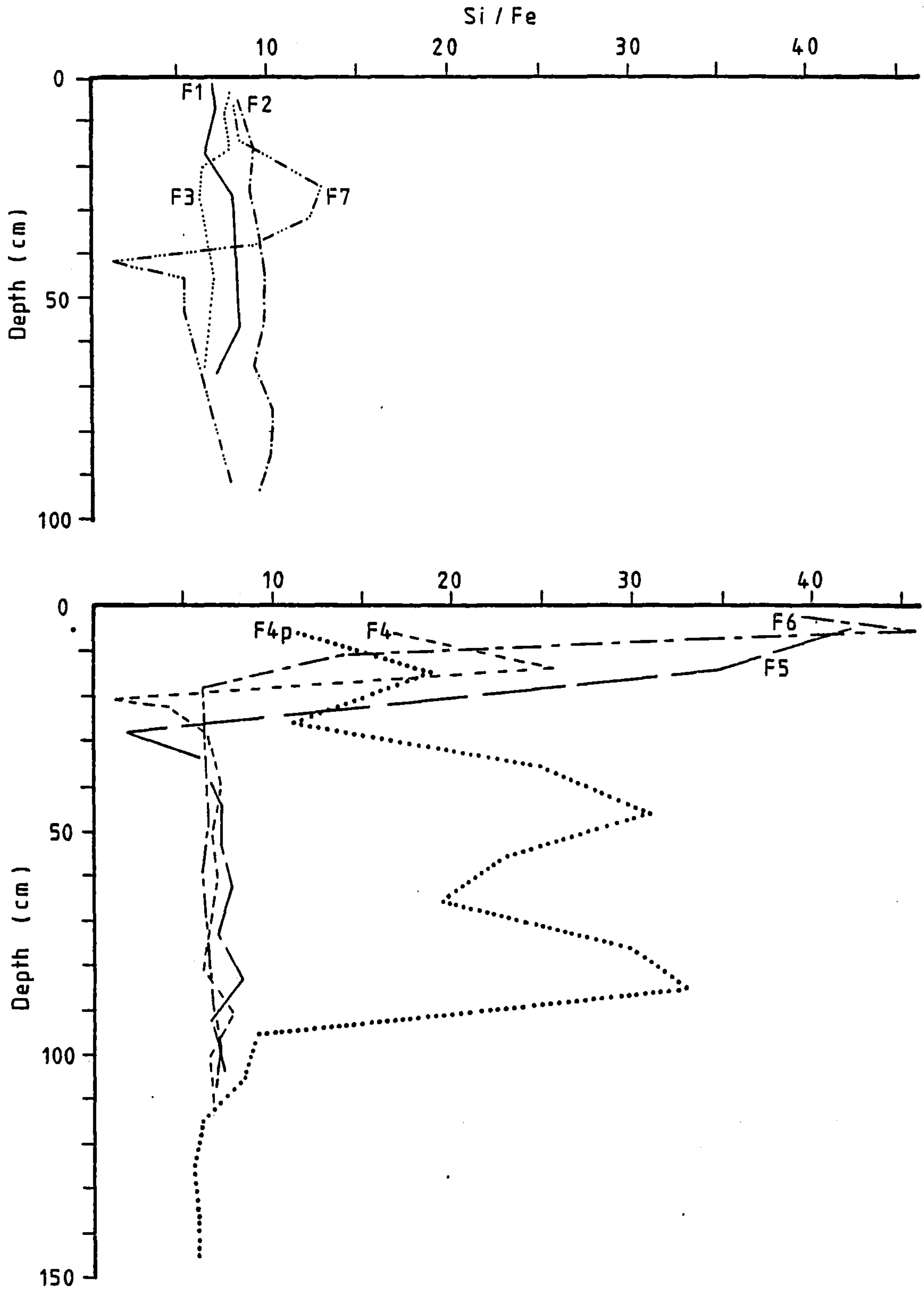


Fig. 67(a): Si/Fe ratios - soils on gentle slopes.

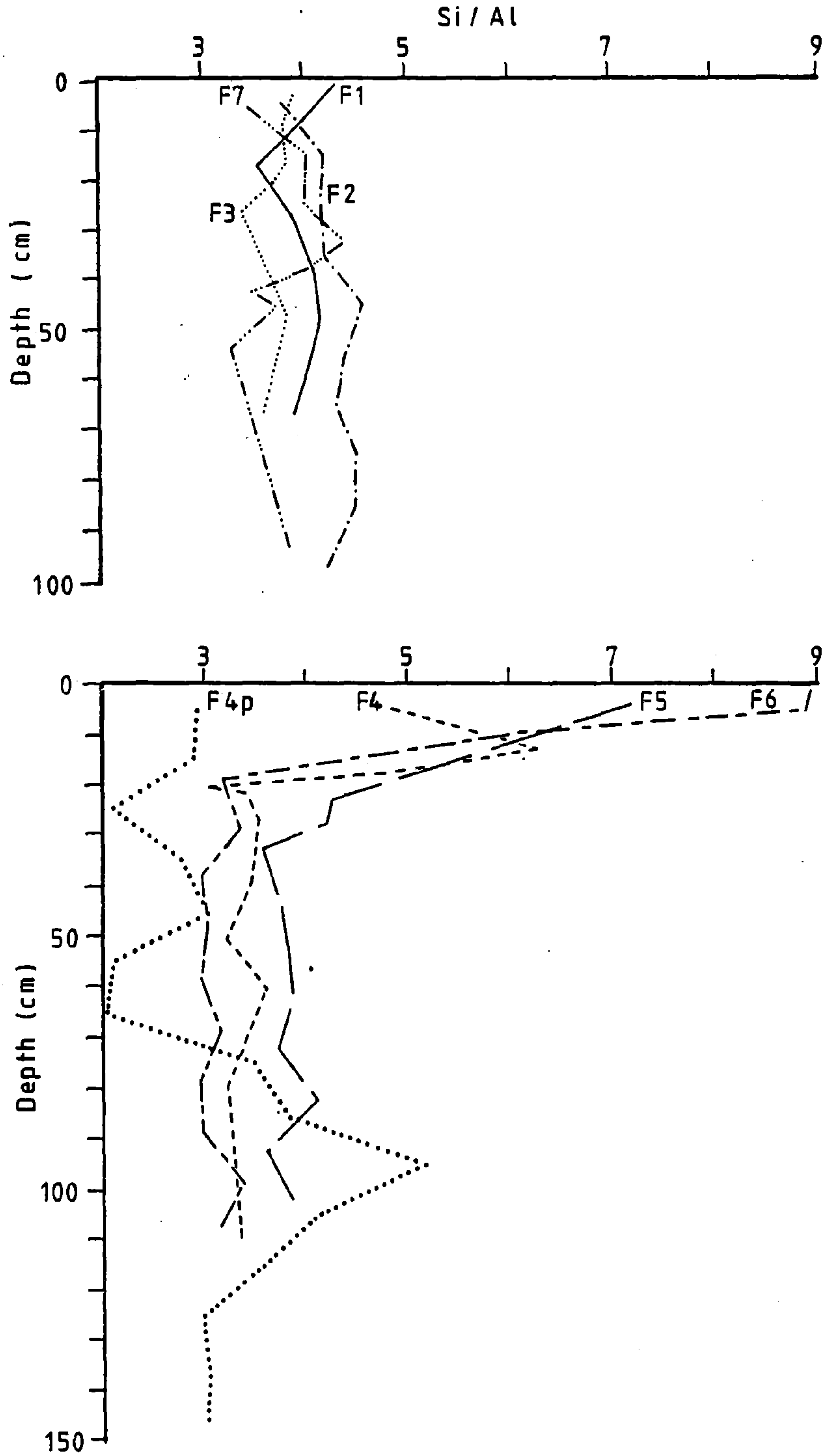


Fig. 67(b): Si/Al ratios - soils on gentle slopes.

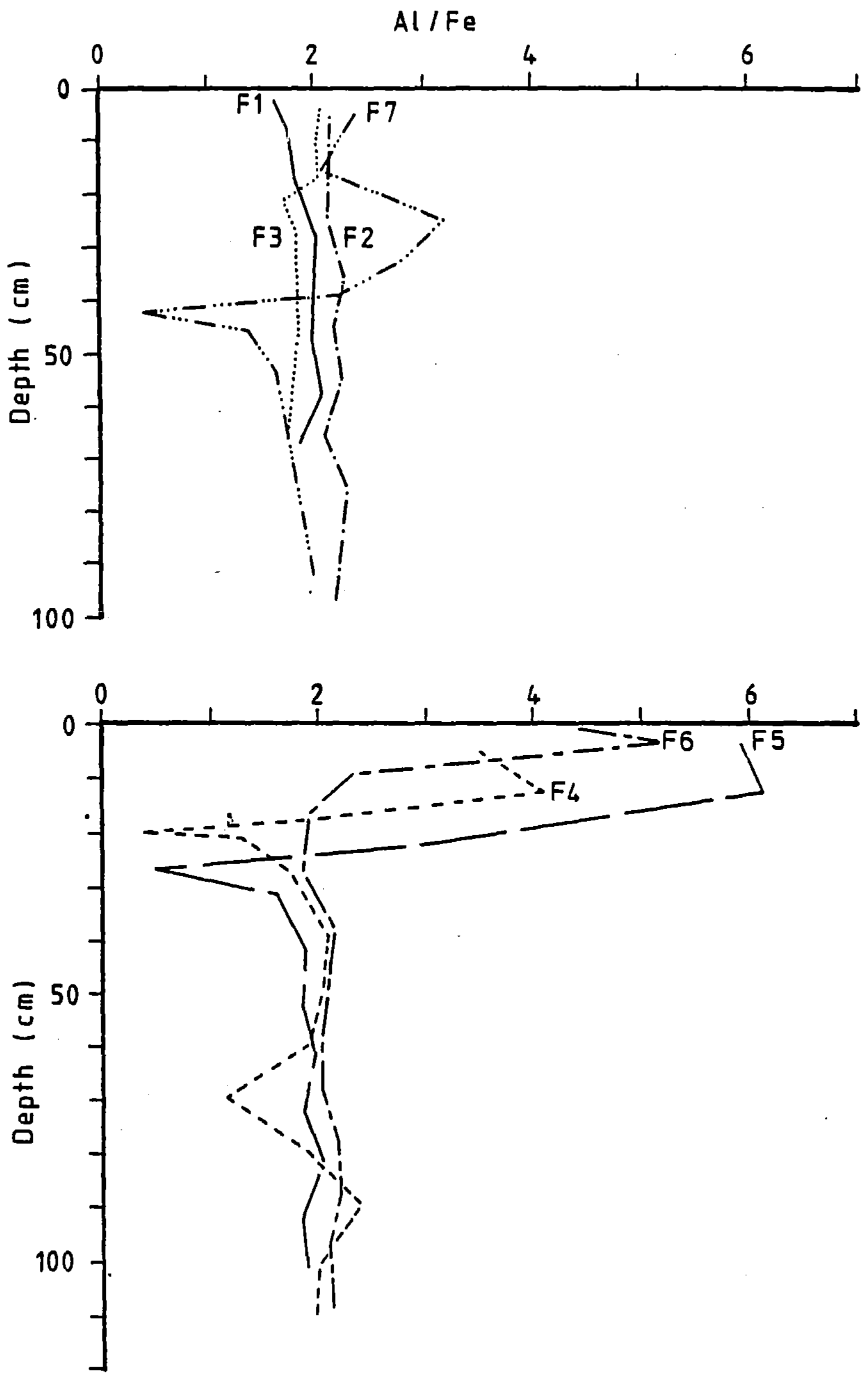


Fig. 67(c): Al/Fe ratios - soils on gentle slopes.

soils and between soils. These ratios showed that the order of mobility of these elements was Fe > Al > Si. There was no evidence that in the young soils Si was being mobilised more rapidly than Fe and Al.

In the recent soils (F1, F2) total Zr concentration (Fig. 68) decreased slightly with depth (270 ppm to 230 ppm) while in F3 the decrease was more marked (from 310 to 230 ppm). In F4, F5 and F6 there was a marked residual concentration in surface horizons (350 to 414 ppm) and minimum concentrations in iron pans. F7 had a similar depth trend but with lower surface horizon concentrations (285-321 ppm). Subsoil concentrations of total Zr showed a very restricted range with most between 210 and 250 ppm. Within the C horizon of individual soils, the range was even more restricted, with most varying by less than 10 ppm.

The total element analyses and elemental ratios supported the assumption of uniformity of soil parent material. Variation in the total element analyses of C horizons was certainly within the range of values obtained from rock analyses (Table 11). Two samples of sandy alluvium from Cropp River and 13 rock samples (8 metapsammites and 5 metapelites), collected from sites throughout upper Cropp basin, were analysed to characterise the range of total element composition in the bedrock. Initially it was envisaged that an average rock composition might be used as the initial material against which the degree of soil development could be evaluated. However, as can be seen from Table 11 there was too much variability in rock composition for this approach to yield meaningful results. This variability was not consistently related to bedrock lithology, except that metapelites had a greater range in elemental composition than metapsammites.

Eluvial-illuvial coefficients (Appendix 3, Fig. 69) eliminate the problem of variability of parent materials

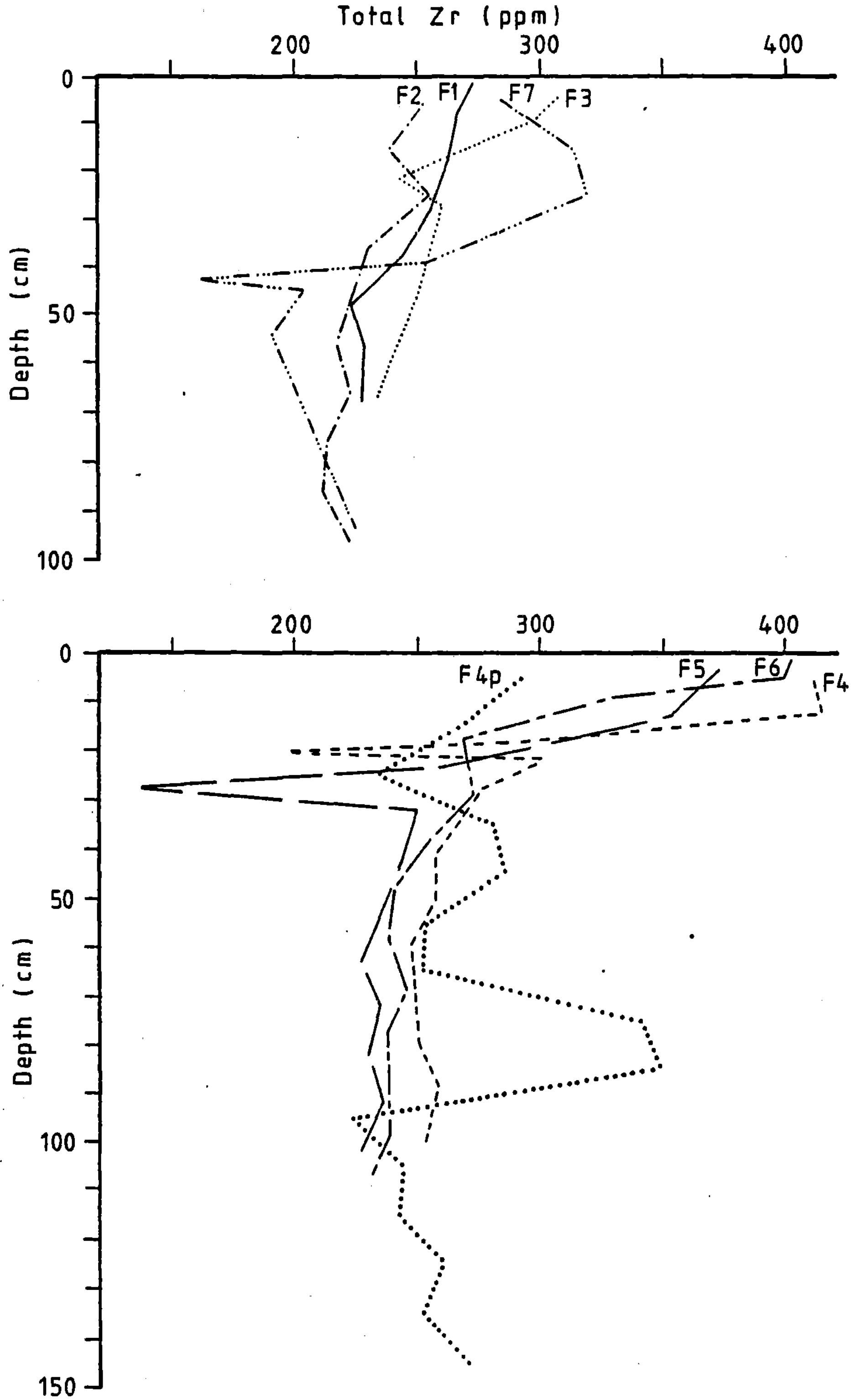


Fig. 68: Total Zr - soils on gentle slopes.

	Ca %	P %	Fe %	Al %	Si %	Zr ppm
	1.54	0.17	4.68	9.42	29.44	185
Meta-	0.74	0.09	4.57	10.00	30.15	175
pelites	0.50	0.05	1.99	6.86	35.34	161
	0.62	0.05	2.35	7.67	34.59	245
	1.23	0.07	3.23	8.50	31.99	248
Range	0.5-1.5	0.05-0.17	2.0-4.7	6.9-10.0	29.4-35.3	161-248
	0.52	0.04	1.80	6.36	36.53	143
	0.88	0.05	2.17	7.14	35.15	191
Meta-	0.97	0.06	3.38	8.34	32.35	205
psam-	1.15	0.05	2.35	7.00	34.95	170
mites	0.53	0.04	1.87	6.40	36.48	194
	0.82	0.04	1.79	6.36	36.30	203
	0.71	0.04	2.09	7.17	34.38	156
	0.71	0.04	1.52	5.83	37.27	243
Range	0.5-1.2	0.04-0.06	1.5-3.4	5.8-8.3	32.4-37.3	143-243
Allu-	0.97	0.08	4.16	7.07	33.28	249
vium	1.06	0.08	3.93	7.37	32.89	246

TABLE 11: Total element analyses of rocks from upper Cropp basin.

between soils by considering element ratios, and the degree of variation of element ratios between the soil horizons and the parent material. Parent material values were approximated by those from the lowermost sample within each soil. For soils on gently sloping sites, variation of C horizon E-I coefficients was generally no more than ± 10%. The majority of coefficients were negative (except in some illuvial horizons) indicating that all elements, including Si, are being lost relative to Zr.

Increasing loss of total Ca from surface horizons with time was indicated by the E-I coefficients of total Ca (Fig. 69a). Losses were least in F1 and F2 (-15 to -30%), greater in F3 (-50%), and greatest in F4 and F5 (-65 to -75%). F6 has similar surface horizon losses to F4 and F5 (-70 to -80%). F7 had relatively small losses of total Ca in surface horizons (-23 to -34%), perhaps as a result of organic cycling in the peaty surface horizons.

E-I coefficients of total P (Fig. 69b) also showed the influence of organic cycling with maximum losses of total P recorded in subsoil horizons rather than surface horizons. There was increasing loss of P in both surface and subsoil horizons with time. Subsoil losses increased in the order: F1 (-10 to -20%), F2 (-30 to -45%), F3 (-55 to -65%), F4 and F5 (-70 to -85%). F6 had similar losses to F4 and F5. F7 also had extremely high subsoil losses (-70 to -90%). The effect of organic cycling of P is clearly shown by $F4_p$ which had large gains of P (+100 to +500%) in the organic surface horizons (Appendix 3).

E-I coefficients of Fe and Al (Figs. 69c, d) showed small losses in F1 and F2 (-10 to -20%), and greater losses in F3 (-20 to -40%). In these soils losses decreased with depth. F4 and F5 both had greater losses (-50 to -80%) in the A_h and E_{ag} horizons, and relative gains in the B_f and B_s horizons. F6 had the greatest loss of all soils in the

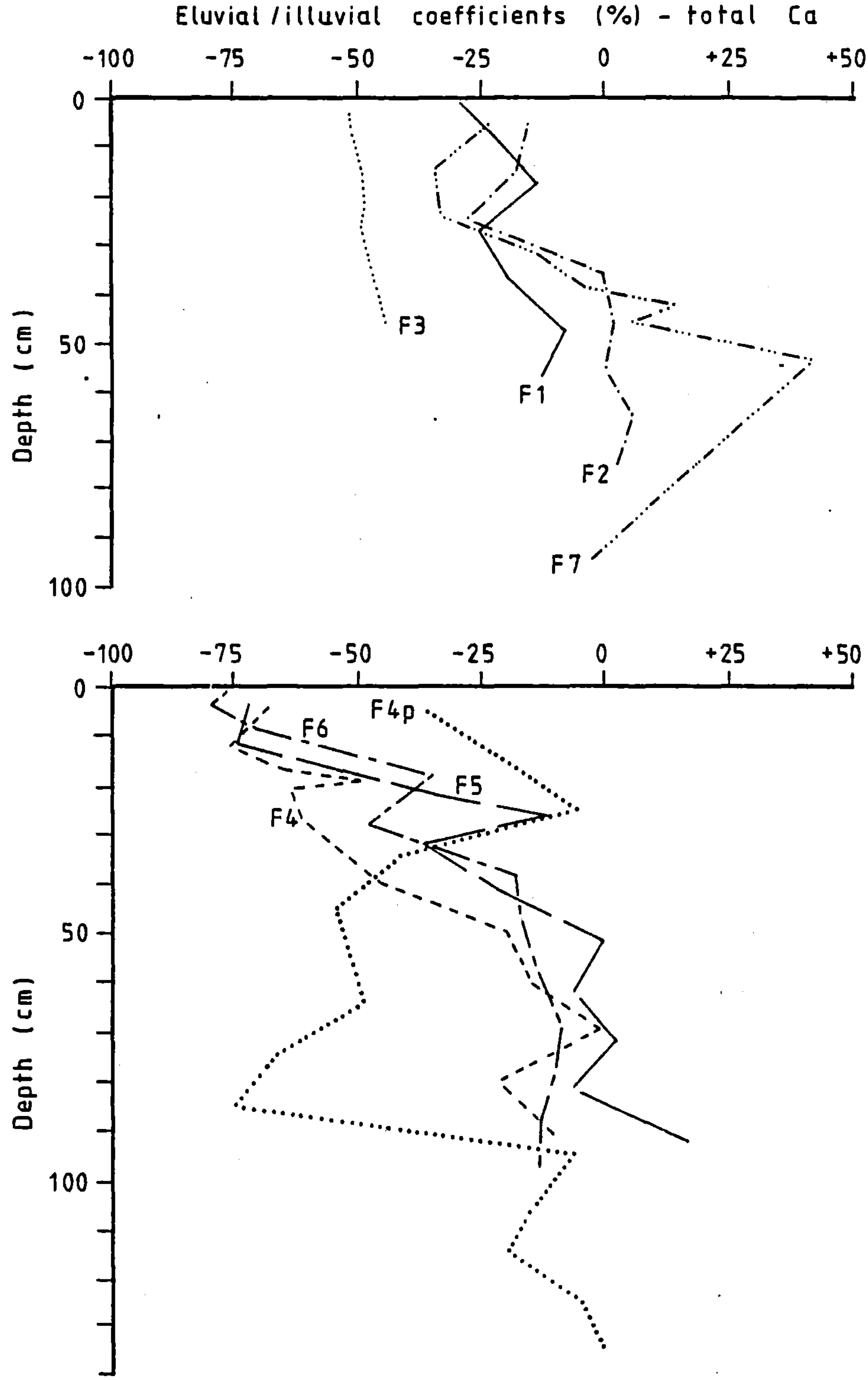


Fig. 69(a): Eluvial-illuvial coefficients of total Ca - soils on gentle slopes.

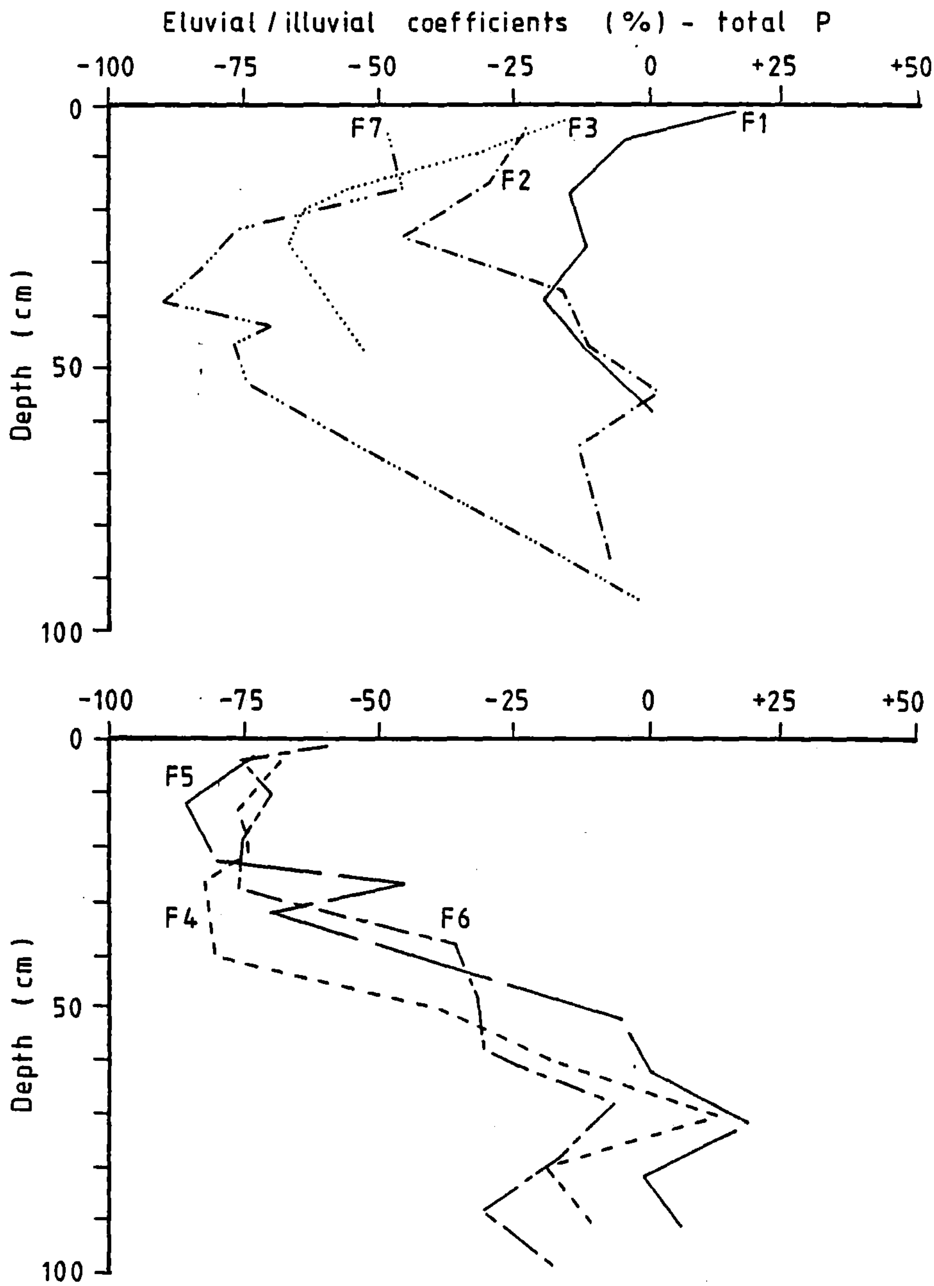


Fig. 69(b): Eluvial-illuvial coefficients of total P - soils on gentle slopes.

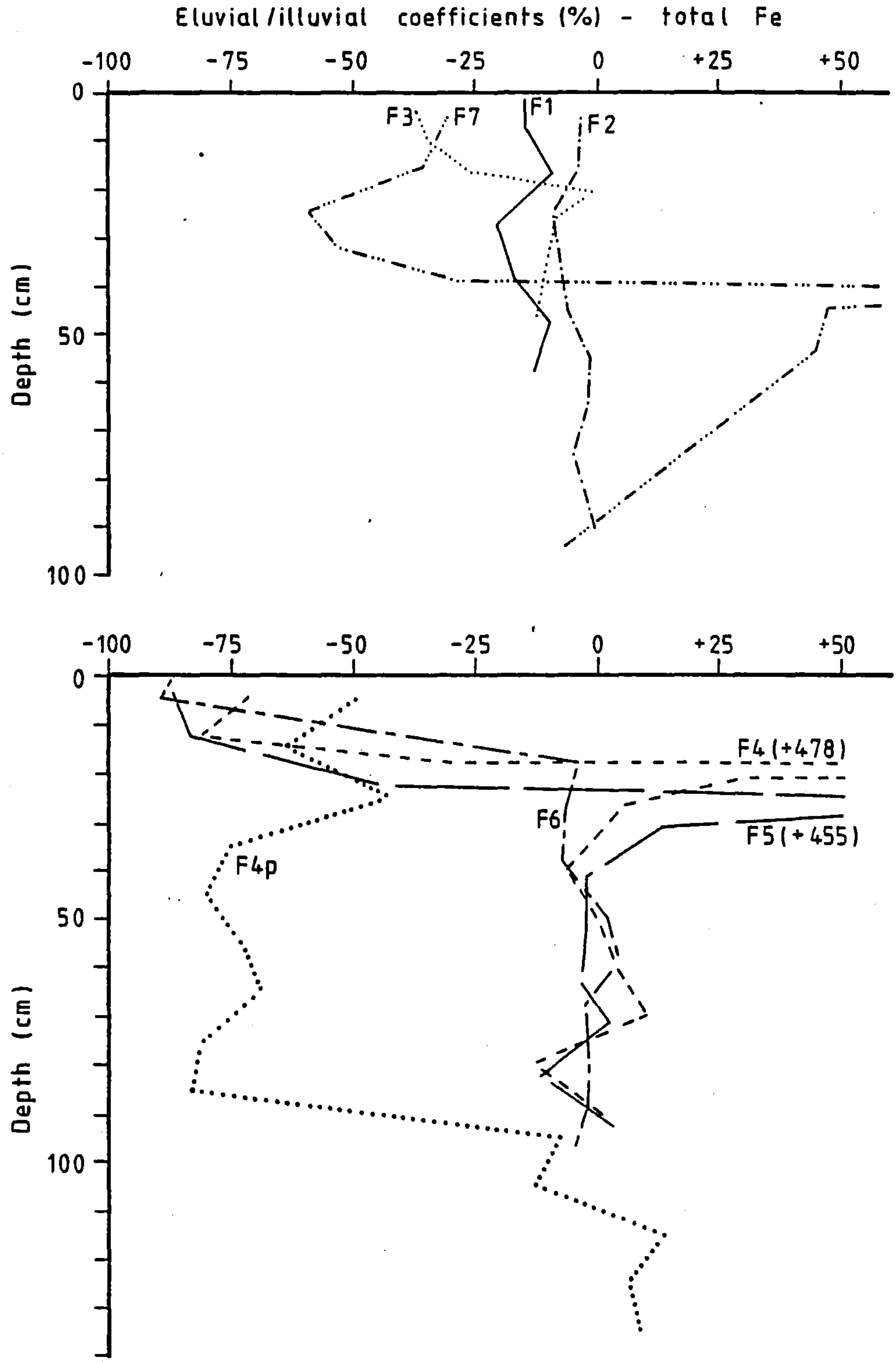


Fig. 69(c): Eluvial-illuvial coefficients of total Fe - soils on gentle slopes.

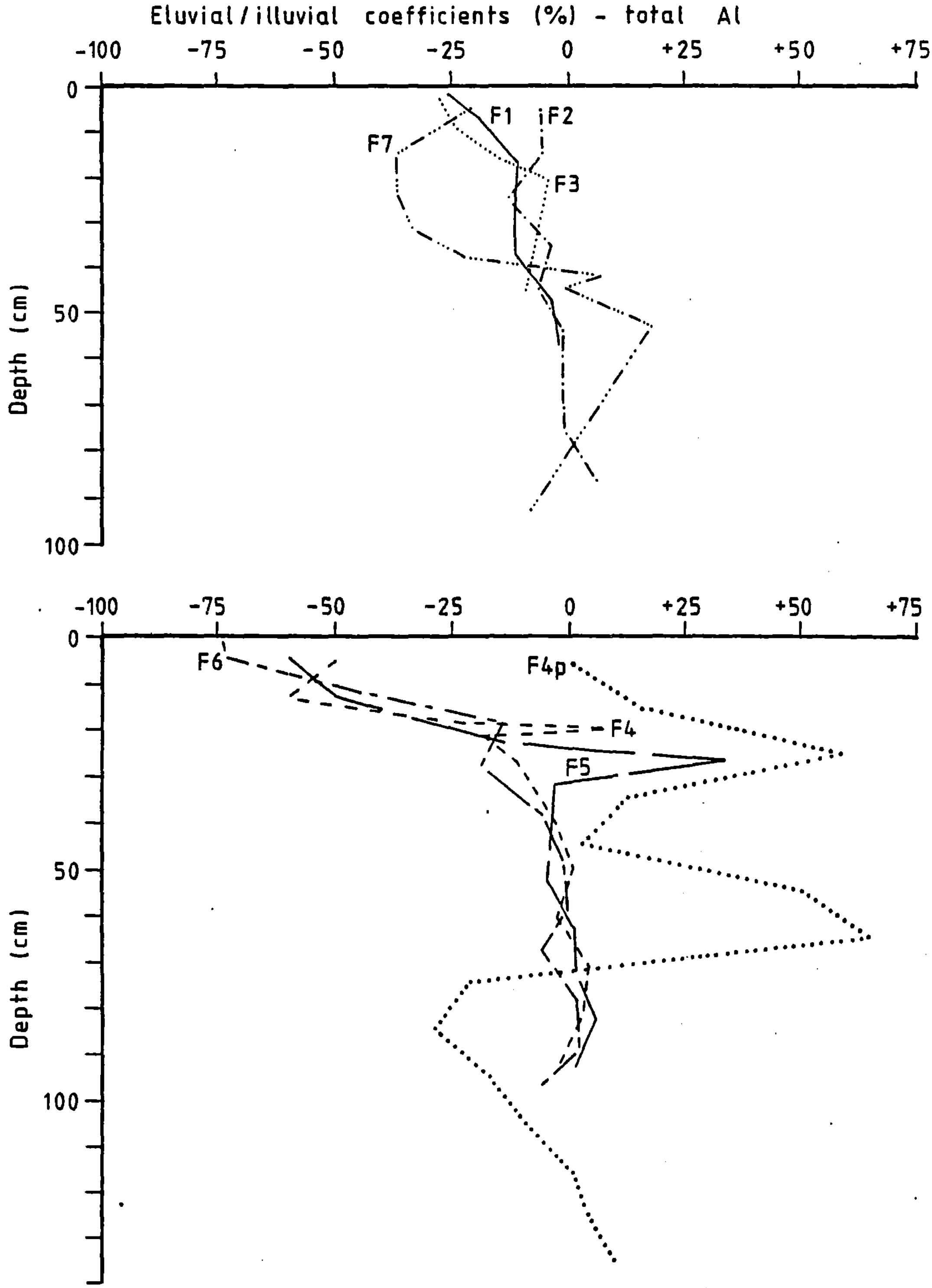


Fig. 69(d): Eluvial-illuvial coefficients of total Al - soils on gentle slopes.

A_h and E_{ag} horizons, but no relative gains in the B_s horizon. F7 had smaller losses of Fe and Al than F4, but greater than F3. $F4_p$ showed large losses of Fe (-50 to -80%) but gains of Al (up to +65%) in the organic surface horizons (Appendix 3).

Losses of Si (Fig. 69e) were clearly of lesser magnitude (maximum of -30%), but negative coefficients for virtually all samples indicated a general loss of Si. Trends in these coefficients were not distinctive, although there was a general tendency for coefficients to become increasingly negative with time. Losses in surface horizons were similar in F1 to F3 (-14 to -20%). In F4 and F5 losses of -25 to -28% were recorded. F6 and F7 also had surface horizon losses of -24 to -28%. The E-I coefficients also showed the relative accumulation of Si in the E_{ag} horizons of F3, F4, F5 and F7.

The E-I coefficients indicate increasing losses of Ca, P, Fe and Al from surface horizons with time, coupled with the redistribution of Fe and Al down the profile. The coefficients generally rank F1 to F5 in order of age, although they did not always clearly distinguish F1 from F2, or F4 from F5. Coefficients calculated for F6 confirmed its interpreted age with losses of Ca, P, Fe and Al comparable to F4 and F5. Losses of Fe and Al in F7 were intermediate between those of F3 and F4.

b) Soils on steep slopes.

Total Ca concentrations in surface horizons had no clear relationship to age for these soils (Fig. 70a). Concentrations declined from S1 to S2 to $S3_{u,l}$, but then increased for S4, $S5_{u,l}$ and S6. The highest recorded values of total Ca were in the peaty surface horizons of $S5_l$ and S6. Nor was there any consistent variation in the soil profile depth trend of total Ca with increasing age. Concentrations in the C horizons ranged from 0.52 to 0.97% total Ca.

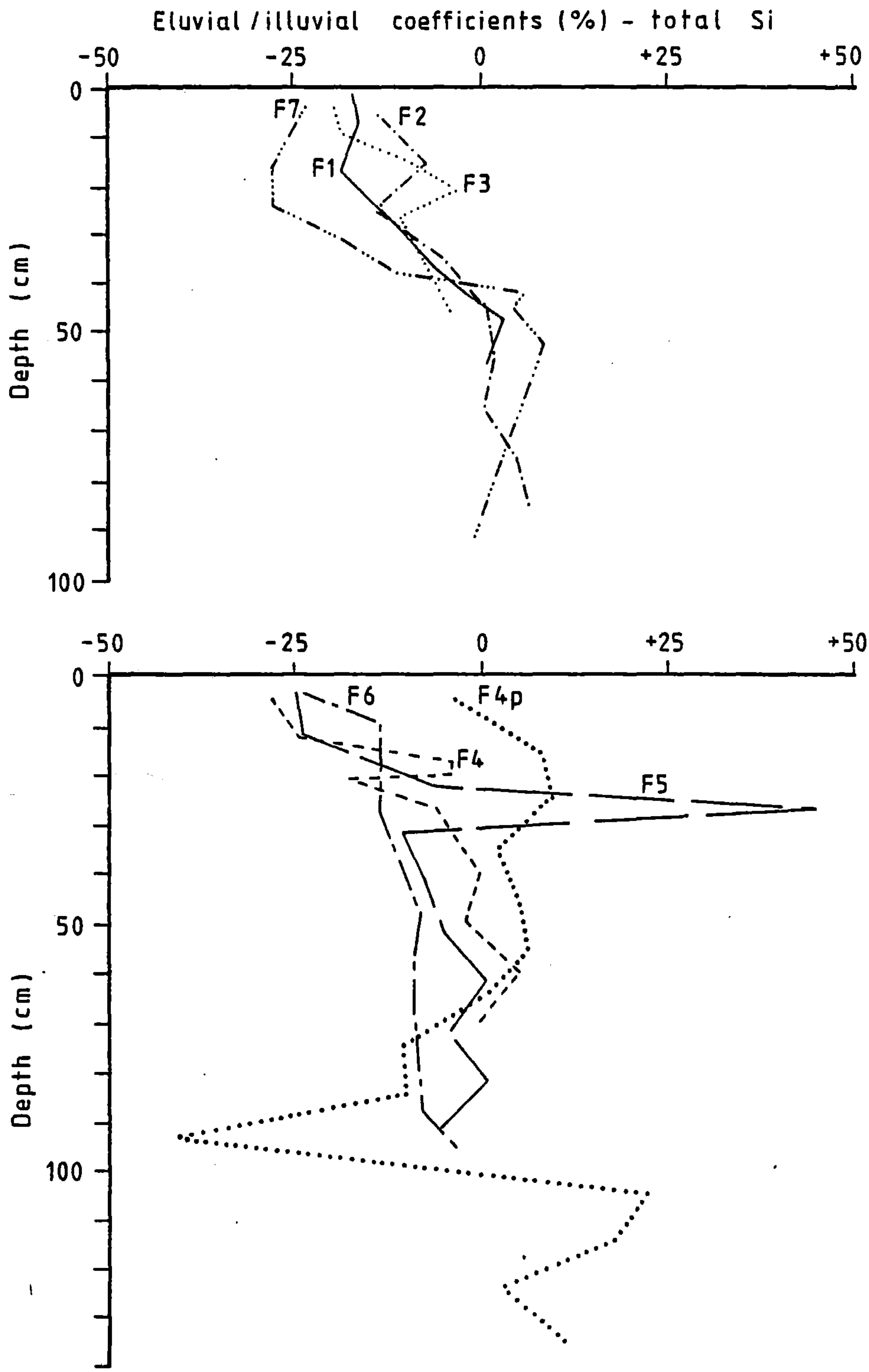


Fig. 69(e): Eluvial-illuvial coefficients of total Si - soils on gentle slopes.

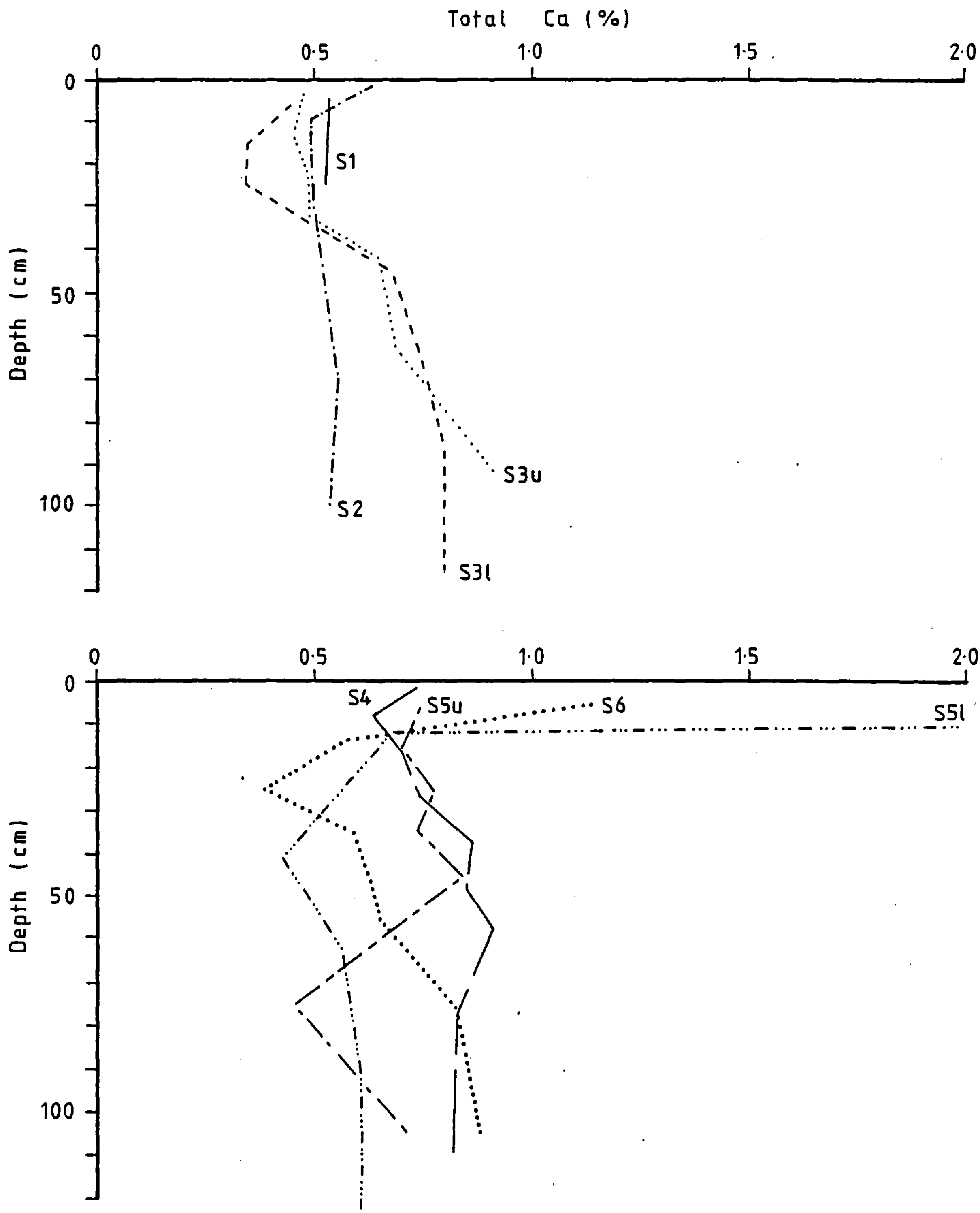


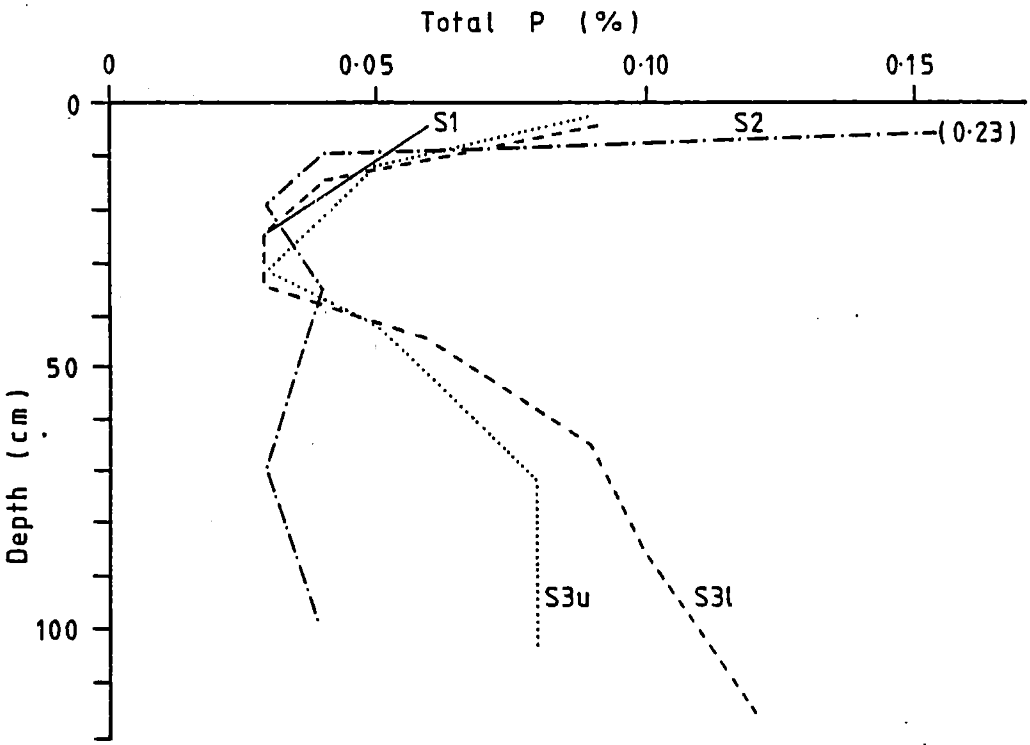
Fig. 70(a): Total Ca - soils on steep slopes.

The distinctive feature of total P analyses was the high concentrations (0.2 to 0.4%) in the peaty surface horizons of S2, $S5_l$ and S6 (Fig. 70b). Other soils had low concentrations (<0.1%), and there were no distinct age related trends. Soil profile depth trend of total P generally showed lowest concentrations in E and B horizons and higher concentrations in surface horizons and C horizons. C horizon concentrations ranged from 0.03 to 0.09%.

Total Fe concentrations (Fig. 70c) were uniformily low in S1 and S2 (2.0 to 2.5%). $S3_u$ and $S3_l$ had their lowest concentrations in the A_h horizon (2.4 to 3.3%). Concentrations increased to a maximum in the B_w (4.8 to 5.3%), and decreased slightly in the C horizon. $S3_l$ had slightly lower values than $S3_u$ throughout. S4 had lower concentrations than $S3_{u,l}$ down to 40 cm, and was similar below 40 cm. $S5_u$ had similar values to $S3_{u,l}$ down to 20 cm and a higher concentration of total Fe in the B horizon (6.3%). $S5_l$ had low concentrations of total Fe in the Bg horizon but similar subsoil values (below 90 cm) to $S5_u$. S6 had the lowest concentration of total Fe in the top 30 cm of the profile (1.0%) of all the soils on steep slopes, and a marked increase to maximum values in the B_s horizon (3.5%), although these maximum values were less than for $S3_{u,l}$ S4 and $S5_u$.

Thus, in A and B horizons of the soil, there was at first an increase in total Fe concentration with increasing age, followed by a decline and redistribution down the profile. The soils were ranked with respect to total Fe concentration and age, except for $S5_u$. Considerable variability in total Fe content of C horizons was evident (2 to 5%), although within the C horizon of each soil values were less variable.

Trends in total Al concentration (Fig. 70d) were similar to those for total Fe, except for the high values of total



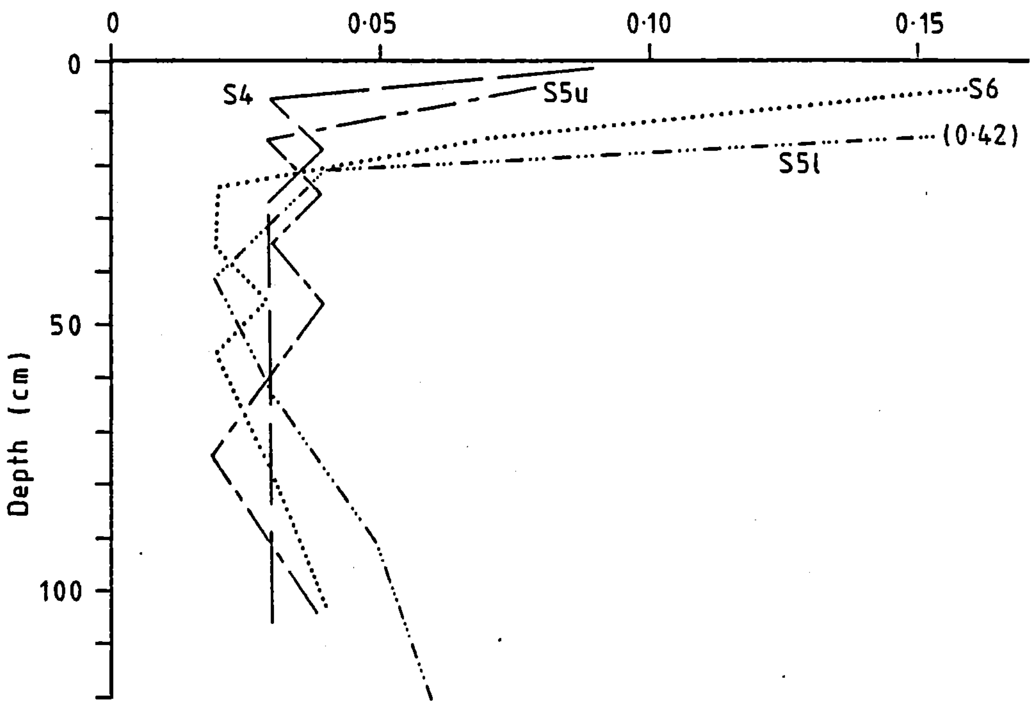


Fig. 70(b): Total P - soils on steep slopes.

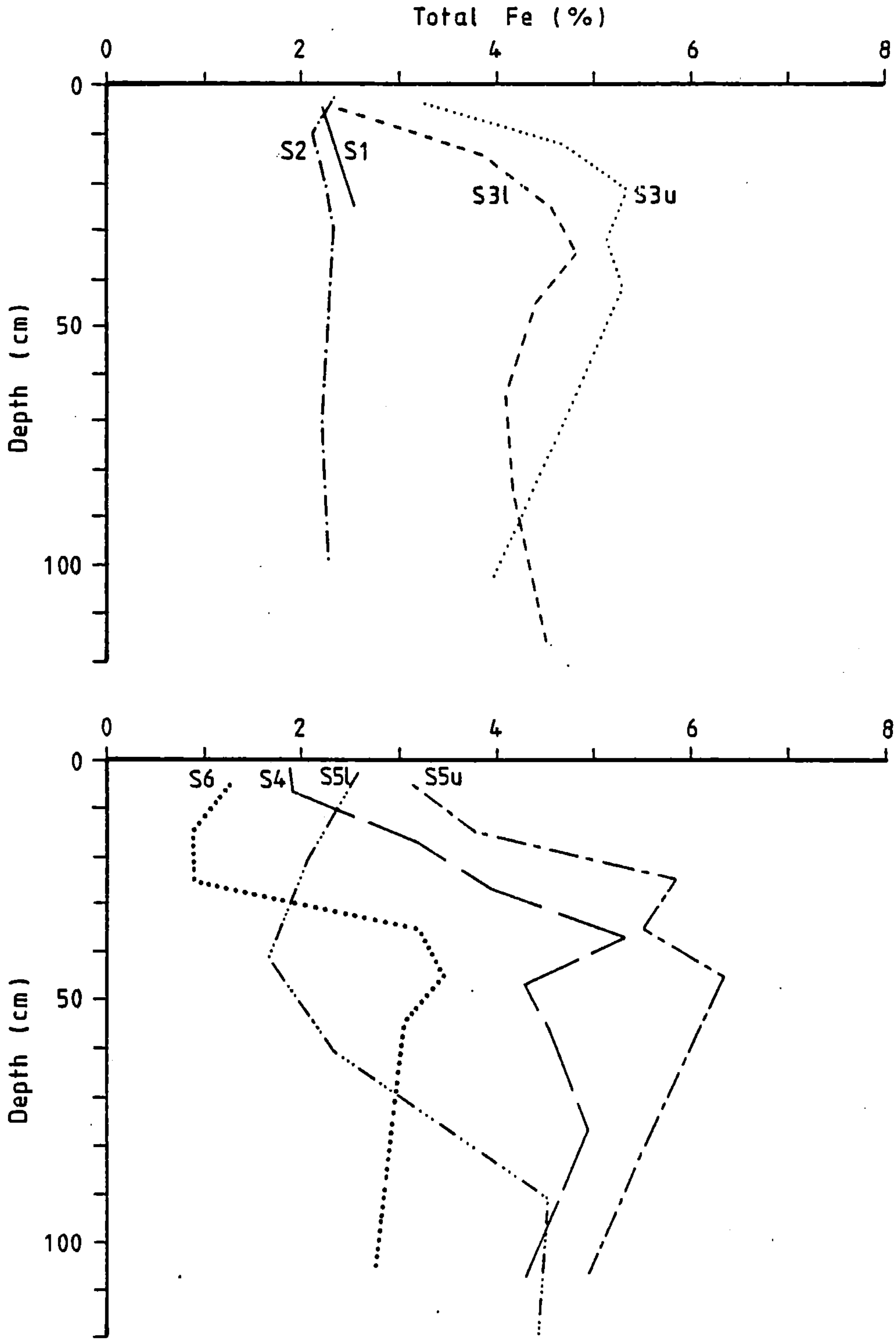
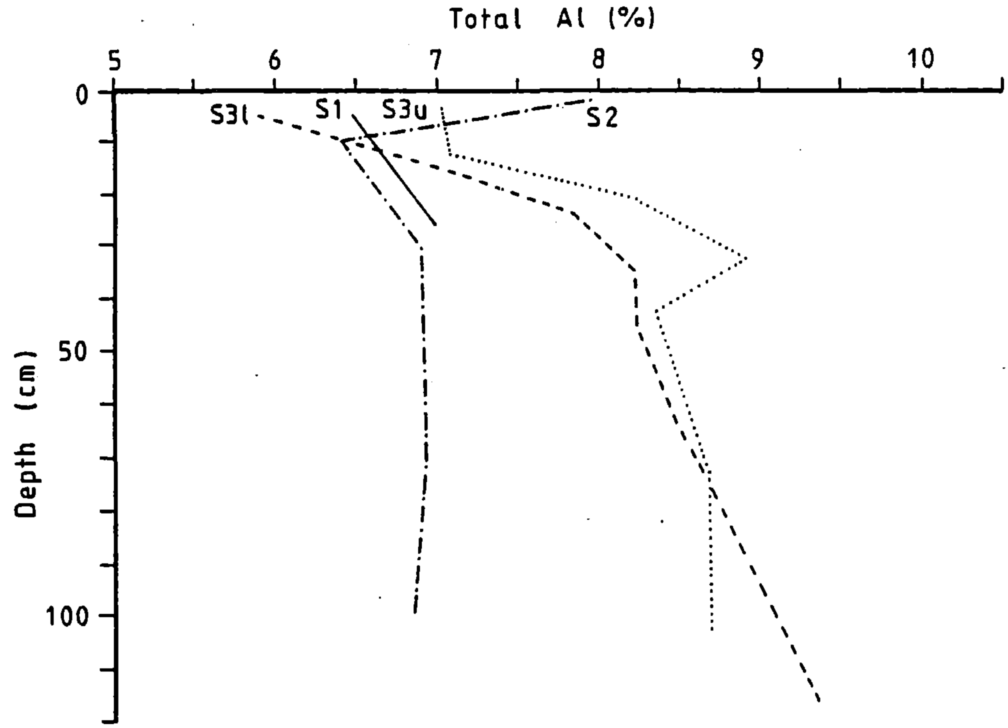


Fig. 70(c): Total Fe - soils on steep slopes.



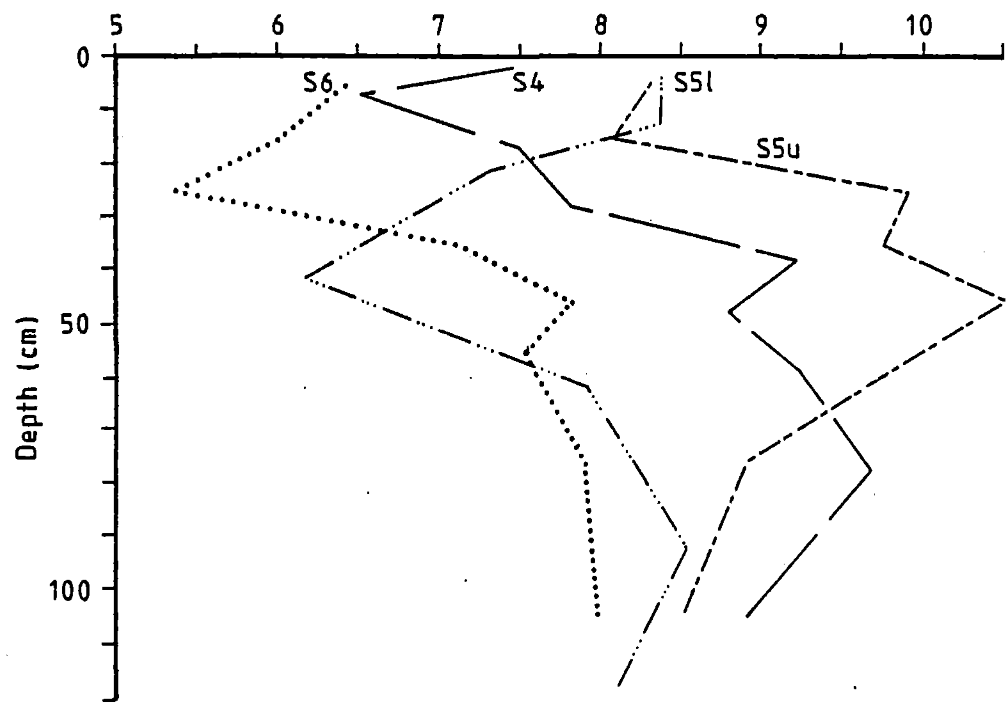


Fig. 70(d): Total Al - soils on steep slopes.

Al in peaty surface horizons (e.g. S2, $S5_l$). S1 and S2 (except for the H horizon) showed little variation in total Al with depth. In $S3_{u,l}$ and S4 there was an increase in total Al with depth, with maximum concentrations in the B_w horizon (8.2 to 9.7%). $S5_u$ had higher concentrations throughout the profile, including the B_s horizon (10.5%), while $S5_l$ had low concentrations of total Al throughout the B_g horizon. S6 showed lower values of total Al throughout the profile, but especially in the A_h and E_{ag} horizons (5.4 to 6.4%). Thus concentration of total Al in the B horizon initially increased with increasing soil development, but then declined in older soils. Again, there was considerable variation in total Al concentrations between C horizons (7-9%), but values within any one C horizon were less variable.

The soil profile depth trend of total Si (Fig. 70e) changed with increasing soil age. S1 and S2 had little variation in total Si with depth. $S3_{u,l}$ and $S5_u$ had highest concentrations in the A_h horizon and a decrease with depth. S4 and S6 both had maximum concentrations in the E_{ag} horizon. Because total Si varied widely between C horizons (30 to 36%) it is difficult to make comparisons between soils, other than in terms of the depth trend. There is no clear ranking of the soils, although S6 had the highest total Si concentration in the A_h and E_{ag} horizons (38%) of all soils on steep slopes. $S5_l$ had a different depth trend of total Si to all other soils, with maximum concentration in the Bg horizon.

Ratios of Si/Fe and Al/Fe (Fig. 71a, c) decreased from the recent soils (S1, S2) to the yellow-brown earths ($S3_u$, $S3_l$) as Fe accumulated in the early stages of soil development. Thereafter values of these ratios tended to increase as Si and Al accumulated relative to Fe. This trend indicated the effect of gleying and mobilisation of Fe^{2+}. A similar trend was noted in the Si/Al ratio (Fig.

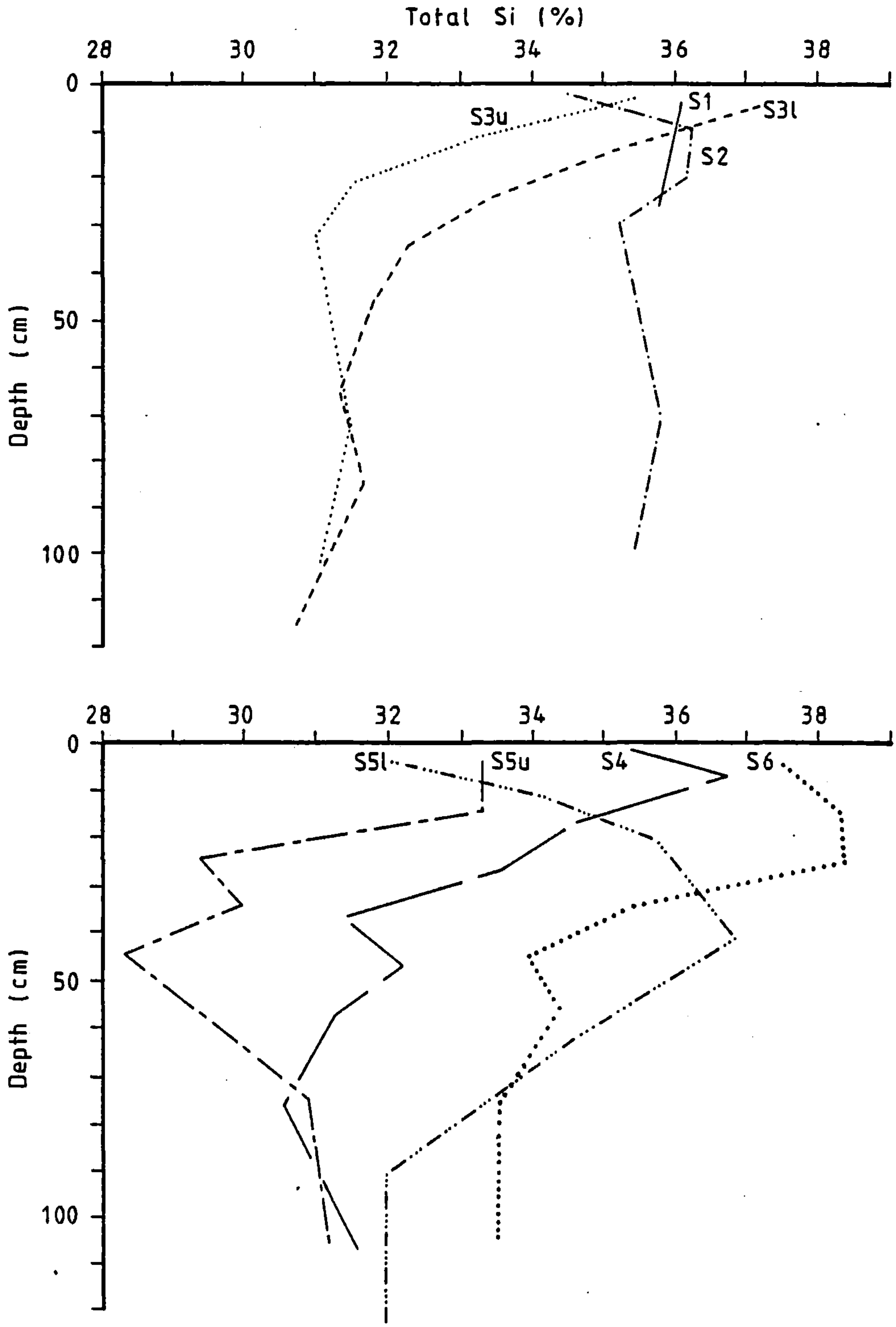
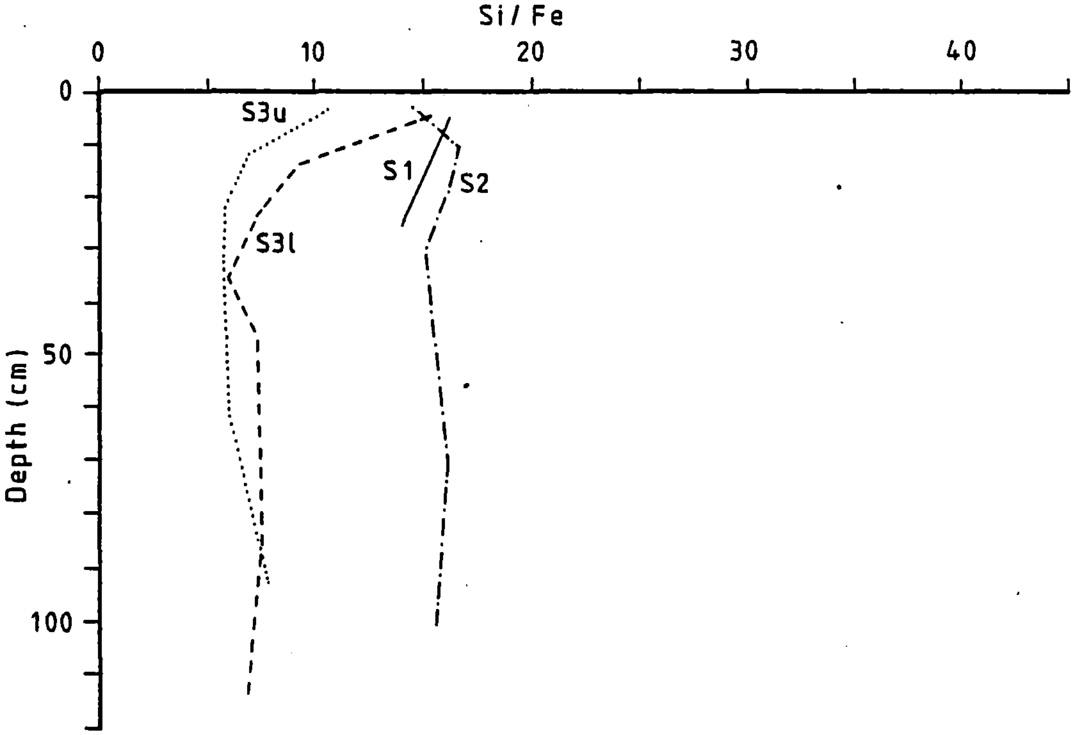


Fig. 70(e): Total Si - soils on steep slopes.



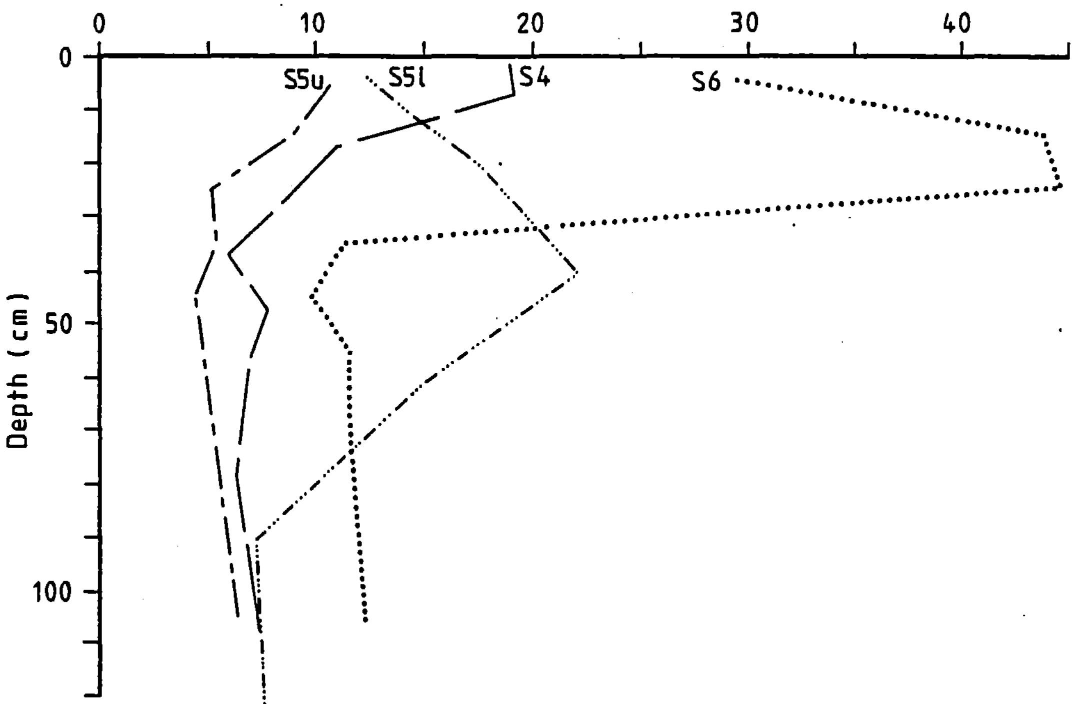


Fig. 71(a): Si/Fe ratios - soils on steep slopes.

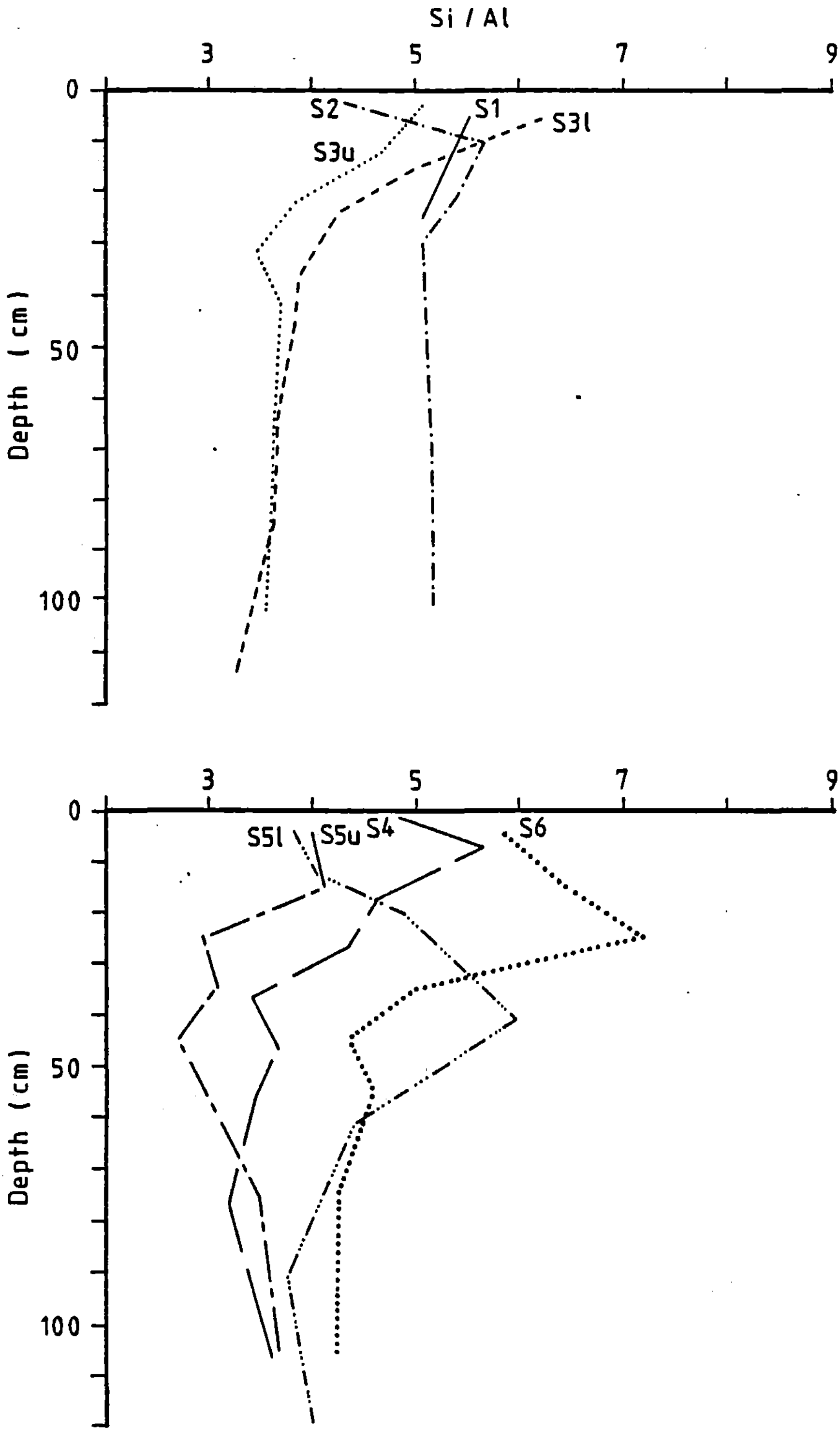


Fig. 71(b): Si/Al ratios - soils on steep slopes.

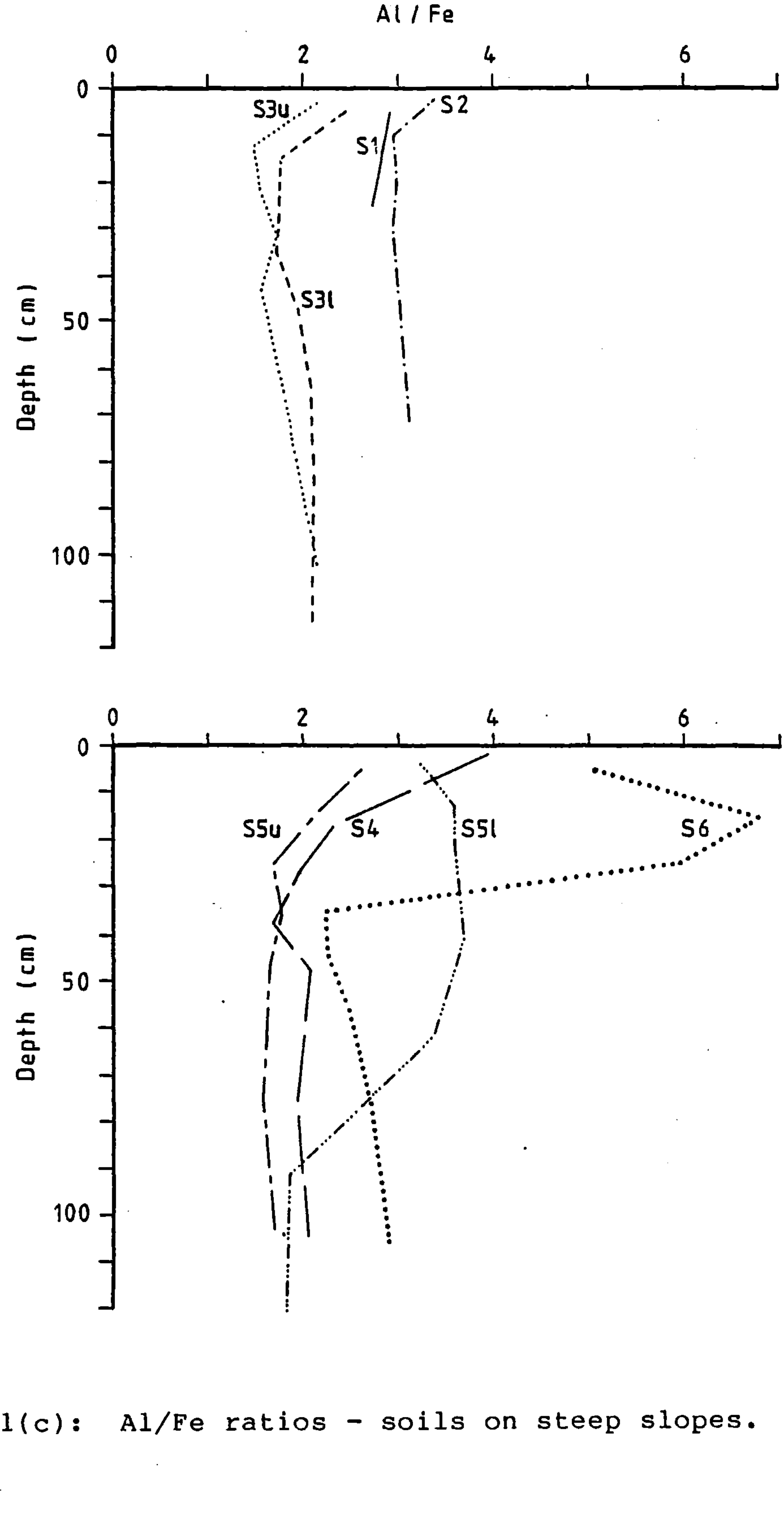


Fig. 71(c): Al/Fe ratios - soils on steep slopes.

71b), i.e. an initial decrease in this ratio followed by an increase. This indicated the effect of podzolisation as pH (H_2O) in the topsoil fell below 4.5.

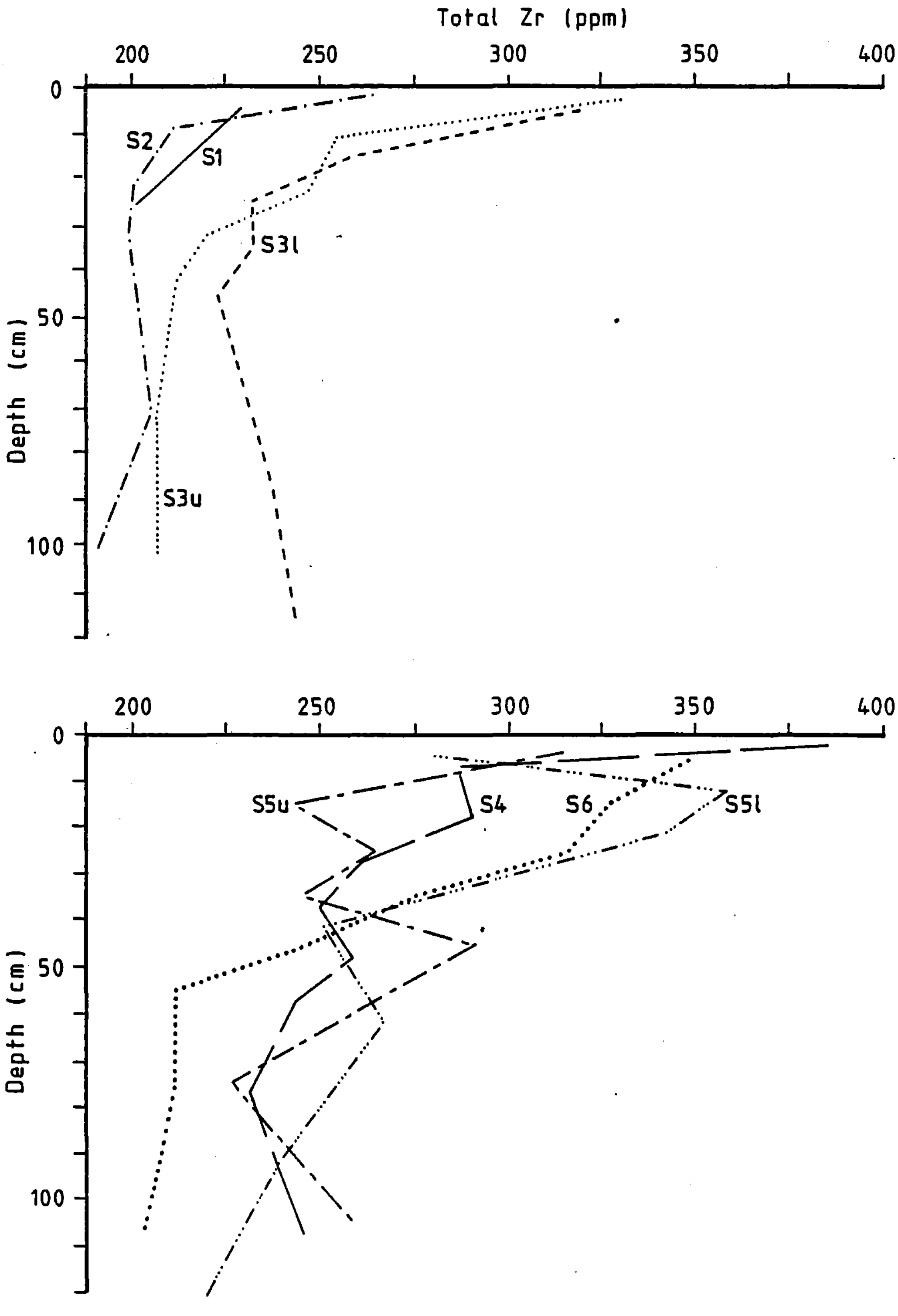
There was a wide range in the values of these ratios between the C horizons of different soils (Si/Al - 3.5 to 5, Si/Fe - 5 to 15, Al/Fe - 3.5 to 5), although within each soil C horizon values were more uniform. This indicated greater variation in the initial chemical composition of soils on steep slopes, and compared with the narrow range recorded for the soils on gentle slopes.

Total Zr concentration (Fig. 72) decreased slightly with depth in S1 and S2, and more markedly in the older soils, due to a residual concentration of Zr in the older surface horizons. C horizon concentrations ranged from 190 to 260 ppm which was within the range of values recorded for bedrock samples, but exceeded that for the soils on gently sloping sites. Within each soil, C horizon concentrations were less variable, generally varying by no more than 30 ppm.

E-I coefficients calculated for the soils on steep slopes were dominantly negative, although coefficients for Ca and P were positive in some peaty surface horizons, and those for Fe, Al and Si were positive in some B horizons. No coefficients are quoted for S1 since insufficient samples were obtained from this soil.

E-I coefficients of total Ca (Fig. 73a) in surface horizons were least in S2 (-10 to -20%) and greatest in S6 (-60% to -70%). $S3_u$ and $S3_l$ had similarly high losses of Ca (-55 to -65%) while S4, $S5_u$ and $S5_l$ had lower losses and were not clearly ranked with respect to age.

Coefficients calculated for total P (Fig. 73b) were highly positive in the peaty surface horizons of S2, $S5_l$ and S6 (up to +345%). All the soils had greatest losses of P in



g. 72: Total Zr - soils on steep slopes.

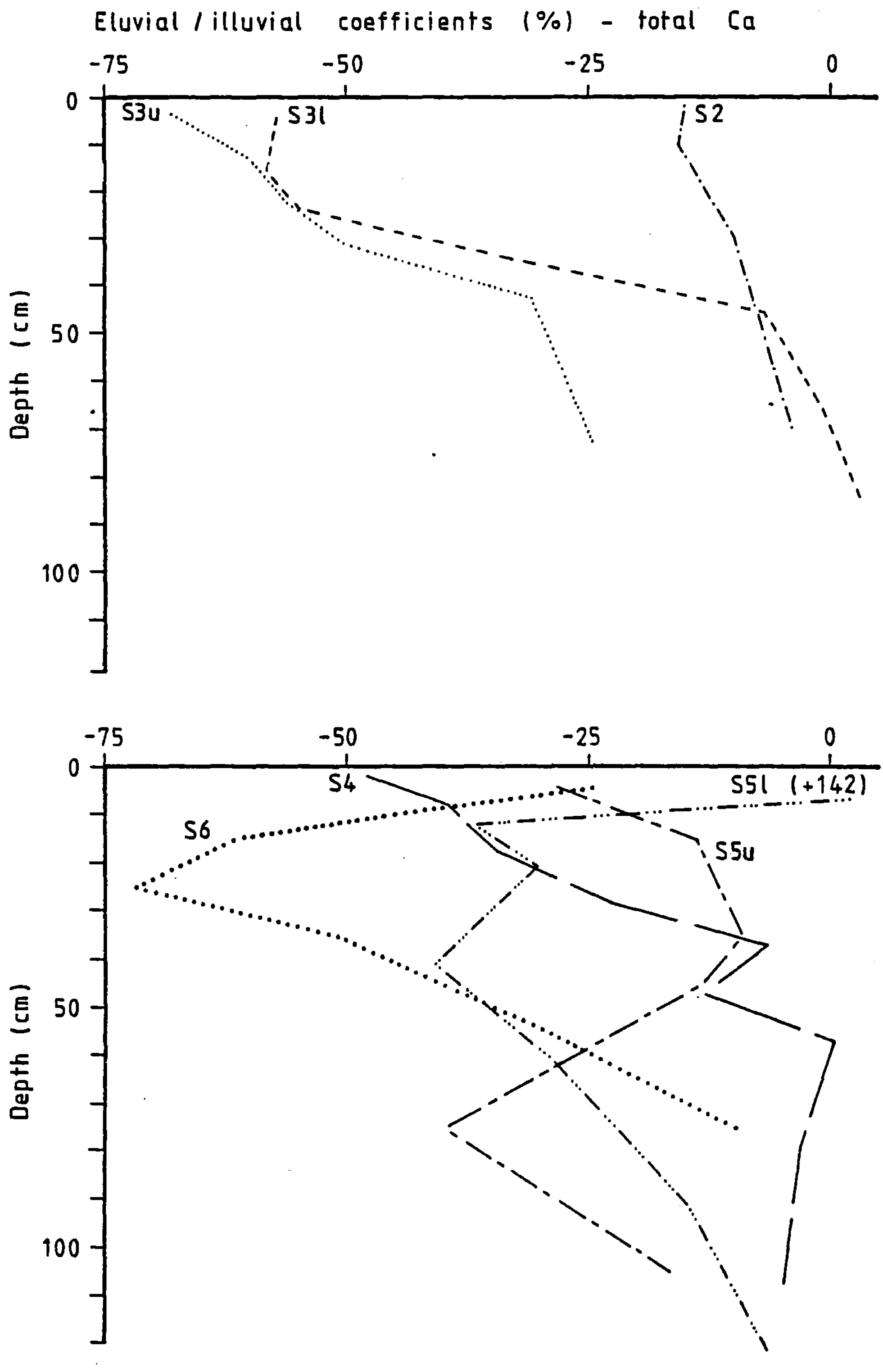


Fig. 73(a): Eluvial-illuvial coefficients of total Ca - soils on steep slopes.

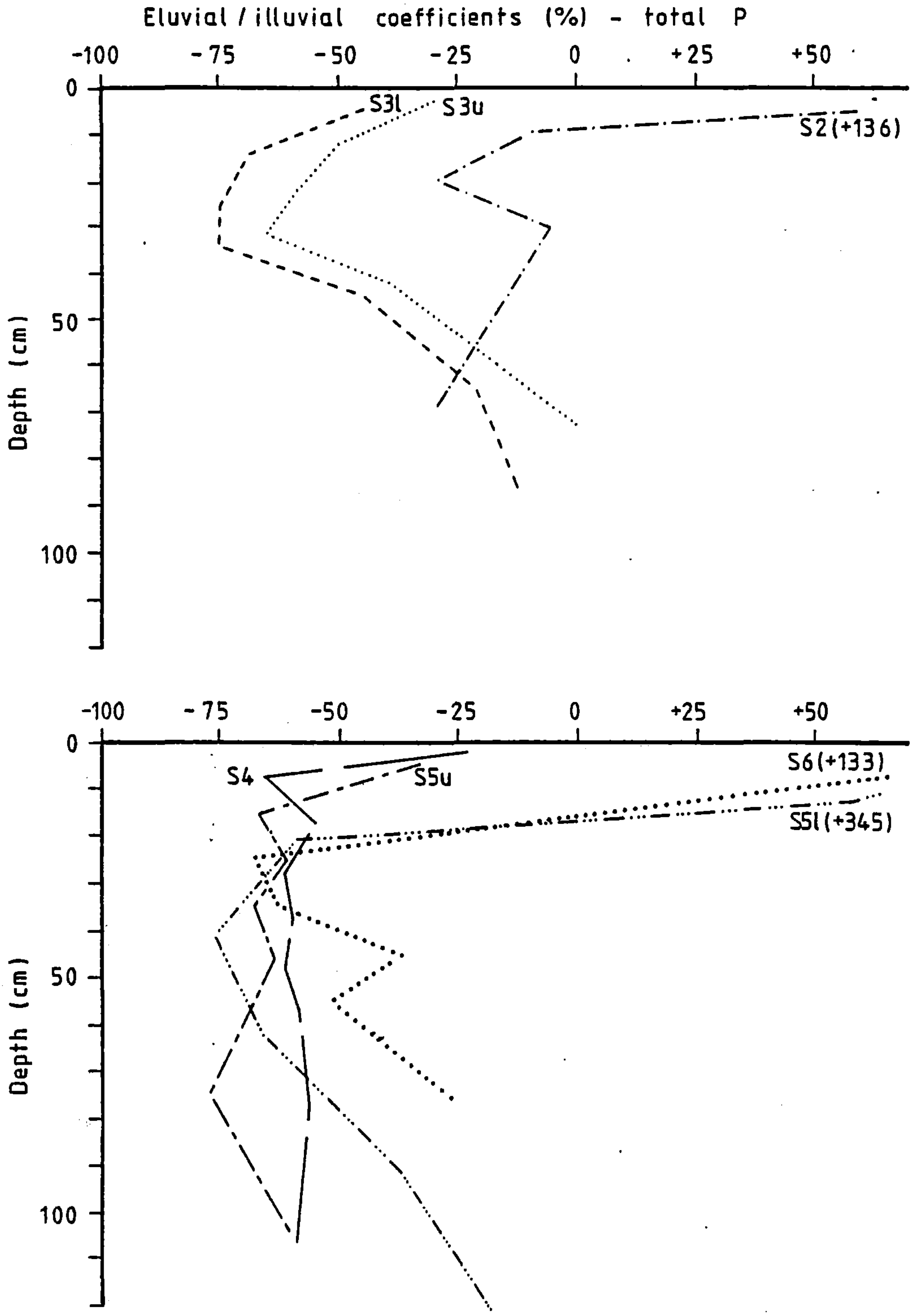


Fig. 73(b): Eluvial-illuvial coefficients of total P - soils on steep slopes.

subsoil (E, B) horizons rather than surface horizons. S2 clearly had lower losses of P (up to -30%) than all the other soils, which showed maximum losses of -50 to -75%.

E-I coefficients of total Fe (Fig. 73c) showed losses from the surface horizons of all soils, while the older soils had gains in the subsoil. Losses from surface horizons increased from -25% in S2, -50 to -60% in $S3_u$ and $S3_l$, -70% in S4, to -80% in S6. $S5_u$ had similar losses to $S3_u$ and $S3_l$, while in $S5_l$ highly negative coefficients (-50 to -60%) persisted throughout the gleyed B_g horizon. In S2, $S3_u$, $S3_l$, S4, and $S5_u$ greatest losses occurred in the A_h and H horizons, but in S6 greatest losses were in both the A_h and E_{ag} horizons. Gains of Fe occurred in the B horizons of $S3_u$, $S3_l$, S4 and $S5_u$. All soils showed similar gains (+10 to 30%). Both pairs of soils on risers showed higher losses of Fe in the soil in the lower slope position.

Coefficients of total Al (Fig. 73d) showed least losses in S2 (-15%) and greatest losses in S6 (-55% in the E_{ag} horizon). $S3_u$ and $S3_l$ appeared to have had greater loss of Al than S4, $S5_u$ and $S5_l$. Small gains occurred in the subsoils of S4 and $S5_l$.

E-I coefficients of total Si (Fig. 73e) had no clear age related trends. S6 had the greatest losses of Si (-25 to -35% in A and E horizons). Losses of Si tended to be less than for Ca, P, Fe and Al, and a number of soils appeared to have gained Si in the subsoil ($S3_l$, S4 and $S5_l$). Both pairs of soils on risers ($S3_u$ and $S3_l$, $S5_u$ and $S5_l$) had markedly different values, with the soil in the upper slope position having the greater loss of Si.

The E-I coefficients indicated a general trend of increasing loss of Ca, P, Fe, Al and Si from surface horizons with time. S2 generally showed the least loss and S6 the greatest. Other soils were not consistently ranked with respect to age.

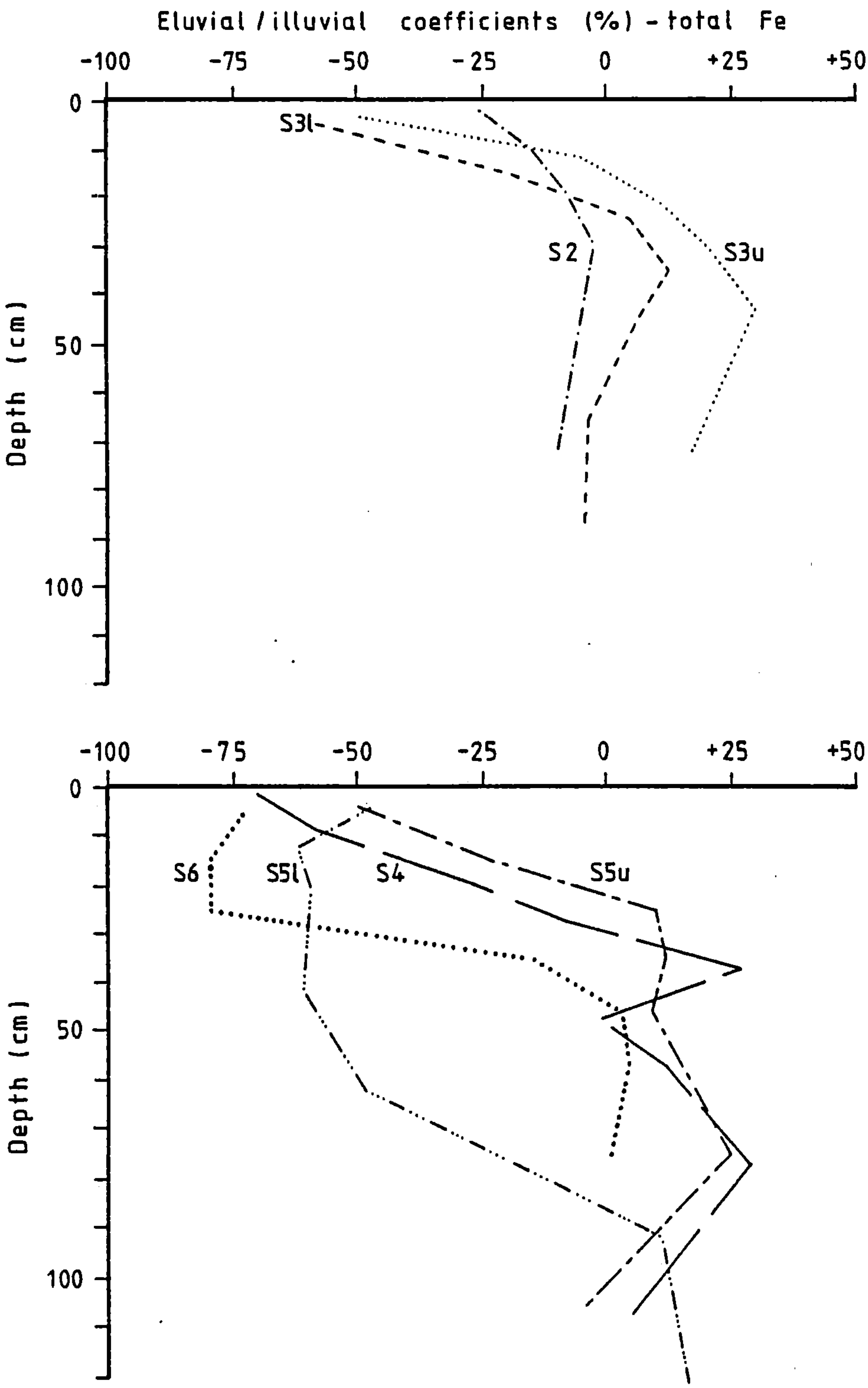


Fig. 73(c): Eluvial-illuvial coefficients of total Fe - soils on steep slopes.

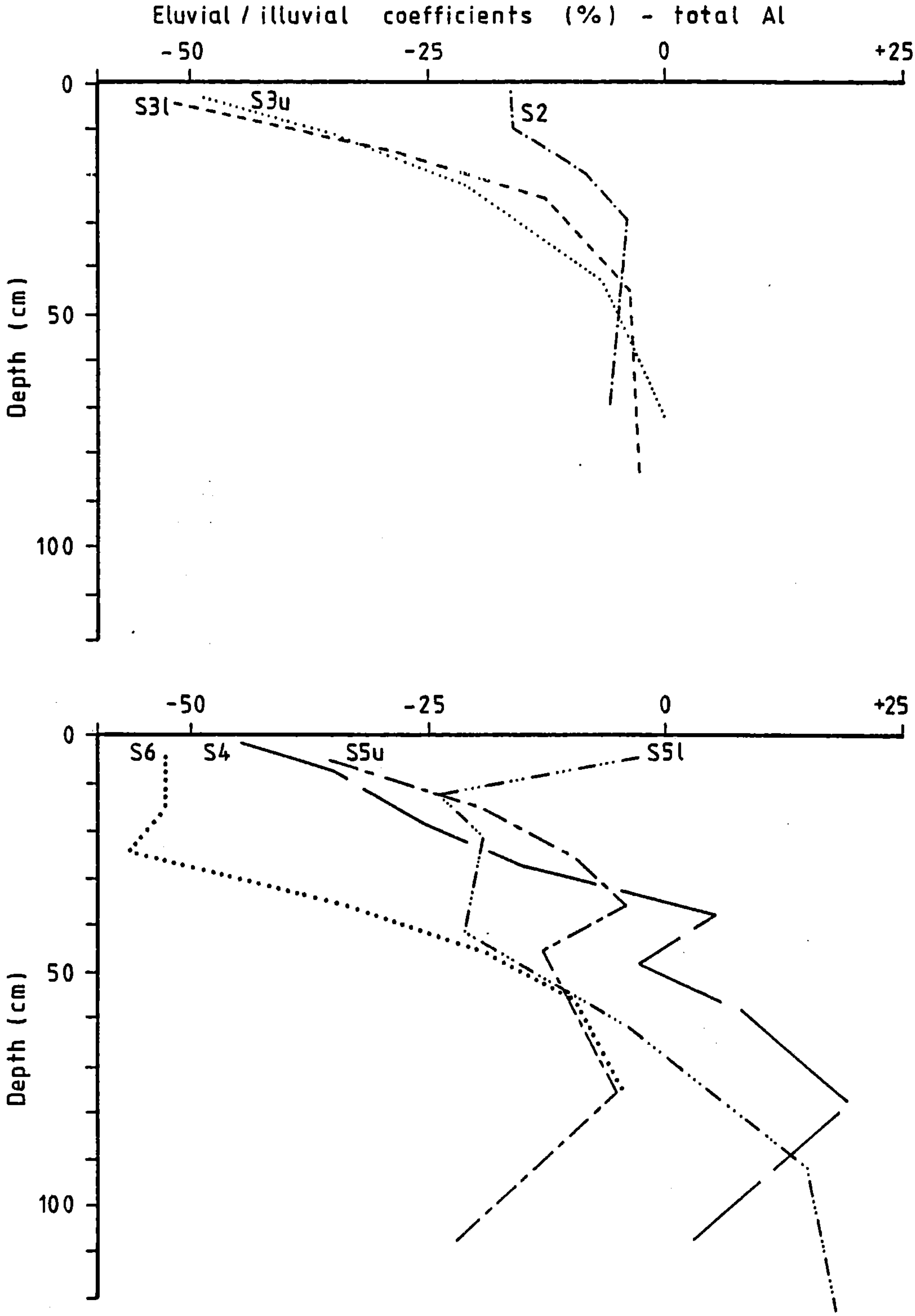


Fig. 73(d): Eluvial-illuvial coefficients of total Al - soils on steep slopes.

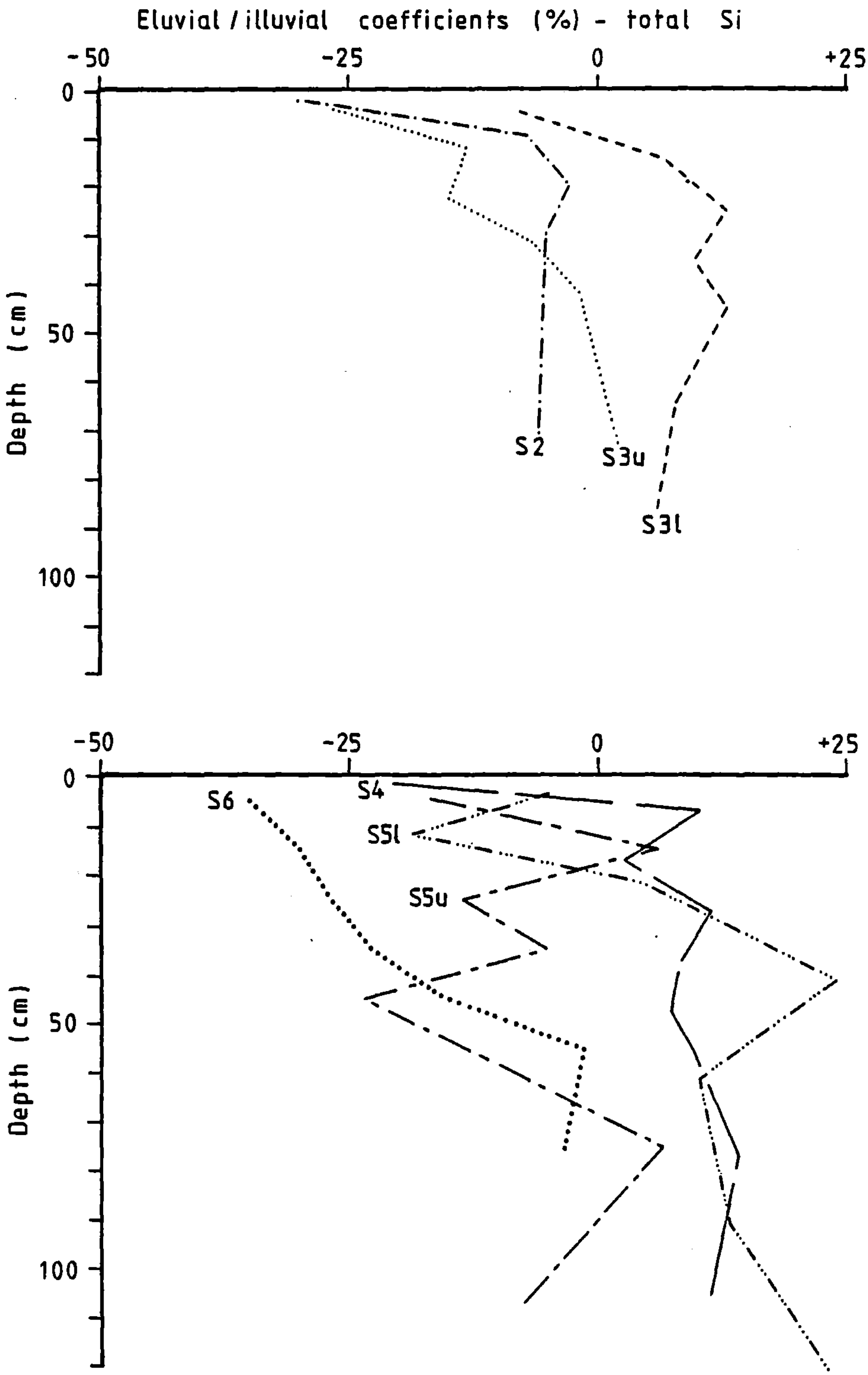


Fig. 73(e): Eluvial-illuvial coefficients of total Si - soils on steep slopes.

6.6.6 Loss-on-ignition

Soil organic matter can be estimated gravimetrically, by ignition at temperatures ranging from 350°C to 850°C (Ball, 1964; Davies, 1974; Campbell, 1975). While some of the weight loss is due to loss of structural water from clay minerals (particularly at higher temperatures), there is a good correlation between measured values of LOI and organic carbon, for a range of ignition temperatures and durations. The method adopted here, used an ignition temperature of 550°C and variable duration depending on organic content. A number of samples were analysed for organic carbon as a check that LOI gave a reasonable estimate of organic matter content. Total C is well correlated with LOI for the 19 analysed samples (see 6.6.7). Values of LOI are tabulated in Appendix 3.

a) Soils on flat sites.
LOI decreased down the profile for all soils without iron pans (i.e. F1, 2, 3, 6). In F4, F5 and F7, an increase in LOI in the iron pan suggested an accumulation of organic matter. There was no distinct relationship between LOI in the A_h horizon and soil age.

b) Soils on steep slopes.
LOI decreased with profile depth for all these soils although subsoil values were variable in some soils (e.g. S4, S5). No distinct relationship between surface horizon LOI and soil age was present, although the older soils tended to have more peaty surface horizons. LOI was generally higher in the surface horizons of these soils than it was in the soils on gentle slopes. In the younger pair of soils ($S3_u$ and $S3_l$) there was no difference in LOI in the surface horizon with slope position. In the older pair of soils ($S5_u$, $S5_l$) there was a major contrast between upper slope (19% LOI) and lower slope position (82% LOI).

6.6.7 Soil fertility analyses

A young soil and old soil from each of the soil sequences

(F2 and F4, S2 and S6) were analysed for organic carbon, total nitrogen, organic and acid-extractable phosphorus, and exchange chemistry. (Tables 12 and 13).

6.6.7.1 Organic matter

Organic C and LOI correlated well (r = 0.99) for the 19 analysed samples:

% C = .63 * LOI -0.85,

Concentrations of organic C were medium (4.7 to 7.8%) in the A_h horizons of F2 and F4 (under tussock grassland), high (14.5%) in S6 and very high (27.0%) in the H horizon of S2 (under shrubby vegetation). Organic C declined with depth in the profile and all subsoils had very low values (<2%).

Surface horizon total N concentrations were low in F2 (0.29%), medium in F4 (0.45%), very high in S2 (1.15%) and high in S6 (0.61%). Concentrations decreased with depth in the profile and all subsoils had low concentrations (<0.3%). Both total C and N were higher under scrub than tussock grassland, probably as a result of greater input of organic matter from shrubs.

C/N ratios were very high in the surface horizons of S2 and S6 (23-25), and high in the more humified topsoils of F2 and F4 (16-17). Thus there was a higher concentration of organic matter under shrubby vegetation, and the organic matter was less decomposed than that accumulating under tussock grassland. C/N ratios declined with depth to very low values in 3 of the 4 profiles examined. In S6 the C/N ratio remained high throughout.

6.6.7.2 Phosphorus

Phosphorus is a key element in the study of pedogenesis, because of the major changes in amounts and forms of P that occur during soil development (e.g. Smeck 1973; Walker and Syers, 1976) and because of its role in determining soil productivity (Walker, 1965). P in organic matter must be supplied almost entirely by the parent material, in unfertilised soils. The redistribution of total P into

Sample Code	Organic C %	LOI %	Total N %	C/N	Phosphorus mg/100 g 0.5M H_2SO_4 soluble (P_a)	Organic (P_0)	Total (P_t)	$\frac{P_a}{P_t}$ %
F2.1	4.7	9.9	0.29	16	9.8	44.7	60	16
.3	1.8	4.6	0.13	15	13.2	27.8	45	29
.10	0.2	0.9	0.03	6	69.2	2.6	76	91
F4.1	7.8	13.8	0.45	17	6.5	32.0	45	14
.2	4.6	8.8	0.31	15	2.9	24.8	34	9
.3	2.6	5.7	0.18	15	2.5	15.3	25	10
.5	1.9	6.0	0.12	16	1.7	13.9	24	7
.12	0.2	1.5	0.03	8	70.1	3.7	72	97
S2.1	27.0	44.9	1.15	23	11.1	96.0	117	9
.2	1.5	3.7	0.09	17	6.4	20.7	35	18
.4	1.1	3.2	0.07	15	8.0	24.0	38	21
.7	0.2	2.1	0.04	6	11.5	11.4	32	36
.11	0.6	1.6	0.03	20	25.3	8.9	39	65
S6.2	14.5	21.8	0.61	24	4.8	40.6	51	9
.3	1.4	2.9	0.06	25	0.5	11.2	20	3
.4	1.3	3.3	0.07	20	2.2	6.8	24	9
.5	1.3	3.4	0.06	21	2.2	15.9	24	9
.7	1.0	3.4	0.05	21	3.7	11.7	23	16
.11	0.6	1.8	0.04	15	22.2	8.5	38	58

TABLE 12: Organic C, total N and phosphorus analyses of selected samples of F2, F4, S2, S6.

organic and various inorganic components with soil development was presented in the discussion of the Franz Josef chronosequence (see Fig. 6).

Trends for H_2SO_4 - soluble P (P_a), organic P (P_o) and total P (P_t) are presented in Fig. 74. P_a is low or very low (<13 mg/100 g) in all but the lowermost sample in each profile. The two older soils (F4 and S6) had lower surface horizon concentrations, which persisted to greater depth in the profile, than the two young soils. Concentration of P_a in the C horizons was far lower for the soils on steep slopes (20 to 25 mg/100g) than for soils on gentle slopes (70 mg/100g). This was due to the differences in parent materials (i.e. weakly weathered alluvium or pre-weathered colluvium), as also shown by pH, oxalate and total element chemistry.

P_o showed the opposite trend to P_a with medium or high concentrations in the A_h and H horizons (32 to 96 mg/100 g), and a decrease with depth to low or very low concentrations (<9 mg/100 g) in the C horizon. P_o was lower in the surface horizons of the two older soils than in the two younger soils. It was also lower in the C horizons of the soils on gentle slopes than in the soils on steep slopes.

P_t was medium or high in the A_h and H horizons (45 to 117 mg/100 g), decreased to low concentrations in the E and B horizons (20 to 35 mg/100 g), and increased in the C horizons. C horizon values were greater in soils from alluvium (>70 mg/100 g in F2 and F4) than from colluvium (<40 mg/100 g in S2 and S6). This reflects parent material differences as discussed previously.

The ratio of P_a/P_t increased with depth in the profile while that of P_o/P_t decreased. The ratio P_a/P_t was far lower for the C horizons formed in colluvium (<65%) than those in alluvium (>90%). The two older soils has lower values of the ratio of P_a/P_t compared to the two younger soils.

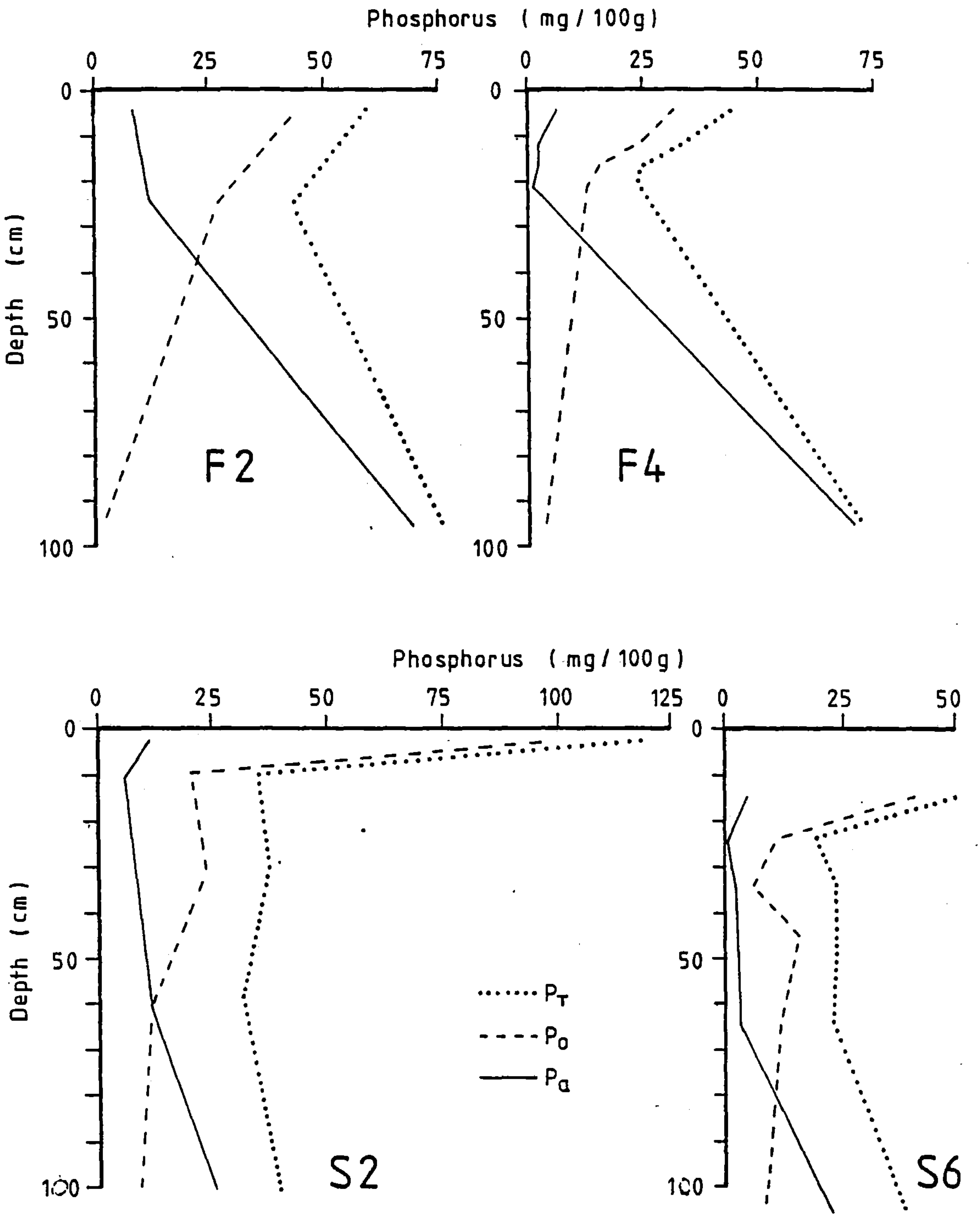


Fig. 74: H_2SO_4- soluble P (P_a), organic P (P_o) and total P (P_t) for selected samples of F2, F4, S2, S6.

The data suggests that P_a and P_t decrease with time. Most P in surface horizons is organically bound and organic P also appears to decrease with time. These trends are consistent with the model of Walker and Syers (1976).

6.6.7.3 Exchange chemistry (Table 13)

Cation exchange capacity was determined by leaching with 1M NH_4OAc, buffered at pH 7. This method overestimates CEC, and underestimates base saturation, in soils with variable charge such as podzols (Gillman and Bell, 1976; Juo *et al*, 1976). Because surface charge, and hence CEC, varies with pH in podzols, the CEC measured by the NH_4 OAc method may bear little relationship to the CEC which exists at the pH of the soil in the field. "Effective Cation Exchange Capacity" (ECEC) was estimated from the sum of exchangeable bases plus KCl-extractable H and Al. This estimate has been shown to agree well with direct estimates of CEC at natural pH (Gillman, 1979; Juo *et al*, 1976).

Highest values of ECEC occurred in surface horizons and were lower for soils on gentle slopes (4.8 to 10.1 m.eq./100 g) compared to soils on steep slopes (10.5 to 12.7 m.eq./100 g). ECEC decreased down the profile except for F4 which had highest values in the E_{ag1} horizon. Values of CEC were generally two to four times those of ECEC. CEC also had maximum values in surface horizons and a decrease with depth. Maximum values were medium (16 to 18 m.eq./100 g) in the surface horizons of soils on gentle slopes, and high or very high (35 to 52 m.eq./100 g) in the surface horizons of the soils on steep slopes.

Concentrations of exchangeable cations were low or very low (<2 m.eq./100 g) in all analysed samples, with little difference between old and young soils. Concentrations decreased down the profile, with very low values in the C horizons of all soils. Exchange sites are primarily occupied by exchangeable Al and H (up to 8 m.eq./100 g). In the young soils (F2 and S2) exchangeable Al and H were at a maximum in

Sample Code	Horizon	Exchangeable cations (m.eq./100g)						TEB[1]	CEC[2]	ECEC[3]	B.S. (%)	
		Ca	Mg	K	Na	H	Al	(m.eq./100g)			pH7[4]	pH MKCl[5]
F2.1	A_{hj}	0.2	0.2	0.18	0.12	0.7	3.4	0.7	16.4	4.8	4	15
.3	2C	0.1	0.1	0.04	0.04	0.4	2.3	0.3	9.1	3.0	3	10
.10	2C	0.2	0.1	0.02	0.03	0.25	0.8	0.35	4.0	1.4	9	25
F4.1	A_h	0.2	0.5	0.24	0.35	2.4	6.3	1.3	18.3	9.0	7	14
.2	E_{ag1}	0.2	0.2	0.17	0.14	1.9	7.5	0.7	14.8	10.1	5	7
.3	E_{ag2}	0.1	0.1	0.10	0.13	1.0	5.2	0.4	12.9	6.6	3	6
.5	$2B_s$	0.1	0.05	0.04	0.04	0.5	3.8	0.2	18.0	4.5	1	4
.12	3C	0.1	<0.05	0.01	0.04	0.1	0.6	0.2	3.8	0.9	5	22
S2.1	H	0.8	0.9	0.69	0.26	2.0	8.0	2.7	52.0	12.7	5	21
.2	C	<0.05	0.07	0.03	0.02	0.3	2.7	0.2	7.7	3.2	2	5
.4	C	0.2	0.07	0.03	0.06	0.3	2.4	0.4	7.4	3.1	5	12
.7	C	<0.05	0.03	0.02	0.01	0.2	1.7	0.1	5.6	2.0	2	5
.11	2C	0.05	0.04	0.02	0.01	0.3	1.8	0.1	6.1	2.2	2	5
S6.2	A_h	2.0	1.2	0.43	0.14	2.9	3.8	3.8	35.3	10.5	11	36
.3	E_{ag}	<0.05	0.06	0.03	0.04	1.1	5.9	0.2	9.1	7.2	2	3
.4	B_s	<0.05	0.05	0.03	0.05	0.9	5.2	0.2	12.1	6.3	2	3
.5	B_s	0.06	0.07	0.03	0.04	1.0	4.3	0.2	13.0	5.5	2	4
.7	BC	<0.05	0.02	0.02	0.02	0.1	2.8	0.1	11.2	3.0	1	4
.11	BC	<0.05	0.03	0.02	0.02	0.1	1.5	0.1	7.6	1.7	1	6

1. TEB = $\sum$Ca + Mg + K + Na

2. CEC = cation exchange capacity determined by ammonium acetate leaching

3. ECEC = "effective cation exchange capacity"; calculated from $\sum$TEB + Al + H

4. Calculated from ratio TEB/CEC

5. Calculated from TEB/ECEC

TABLE 13: Exchange chemistry of selected samples of F2, F4, S2, S6.

the surface horizons, and decreased down the profile. In the old soils exchangeable H concentrations were higher than for young soils and they decreased down the profile. Exchangeable Al was also higher but showed maximum concentrations in the E_{ag} horizons.

Total exchangeable bases were very low (<4 m.eq./100 g) in all samples, including surface horizons, and declined with depth. The surface horizons of S2 and S6 had the highest measured concentrations (2.7 to 3.8 m.eq./100 g), but even these were very low. Base saturation (pH 7) was very low throughout all 4 profiles with highest values generally in surface horizons. Base saturation (at pH of MKCl) yielded higher values (up to 36%) but they were dominantly low.

The results demonstrate the extremely leached nature of these soils, showing little difference between old and young soils except in the greater concentration of exchangeable Al and H in the older soils. Since young soils are as leached as old ones, it appears that elements released by weathering are either assimilated by plants or leached from the soil system. Soil exchange sites are primarily occupied by Al and H.

6.6.8 Particle size analysis

Soil sequences in Westland show changes in particle size distribution within soils of increasing age (Stevens, 1968; Mokma *et al*, 1973; and Campbell, 1975). With increasing soil development and weathering there is an overall decrease in coarse clastic material, and changes in the proportion of sand, silt and clay in the upper soil horizons.

a) Soils on gentle slopes (Fig. 75).
With increasing age of these soils the proportion of silt and clay in surface horizons increased and sand and gravel decreased. Clay content rose from 4% in FO (an alluvial sample) to 10% in the A_h horizons of F1 and F2, and reached 31% in F3. It then declined to 25% in F4 and 19% in F5. This decline may be partly a function of soil

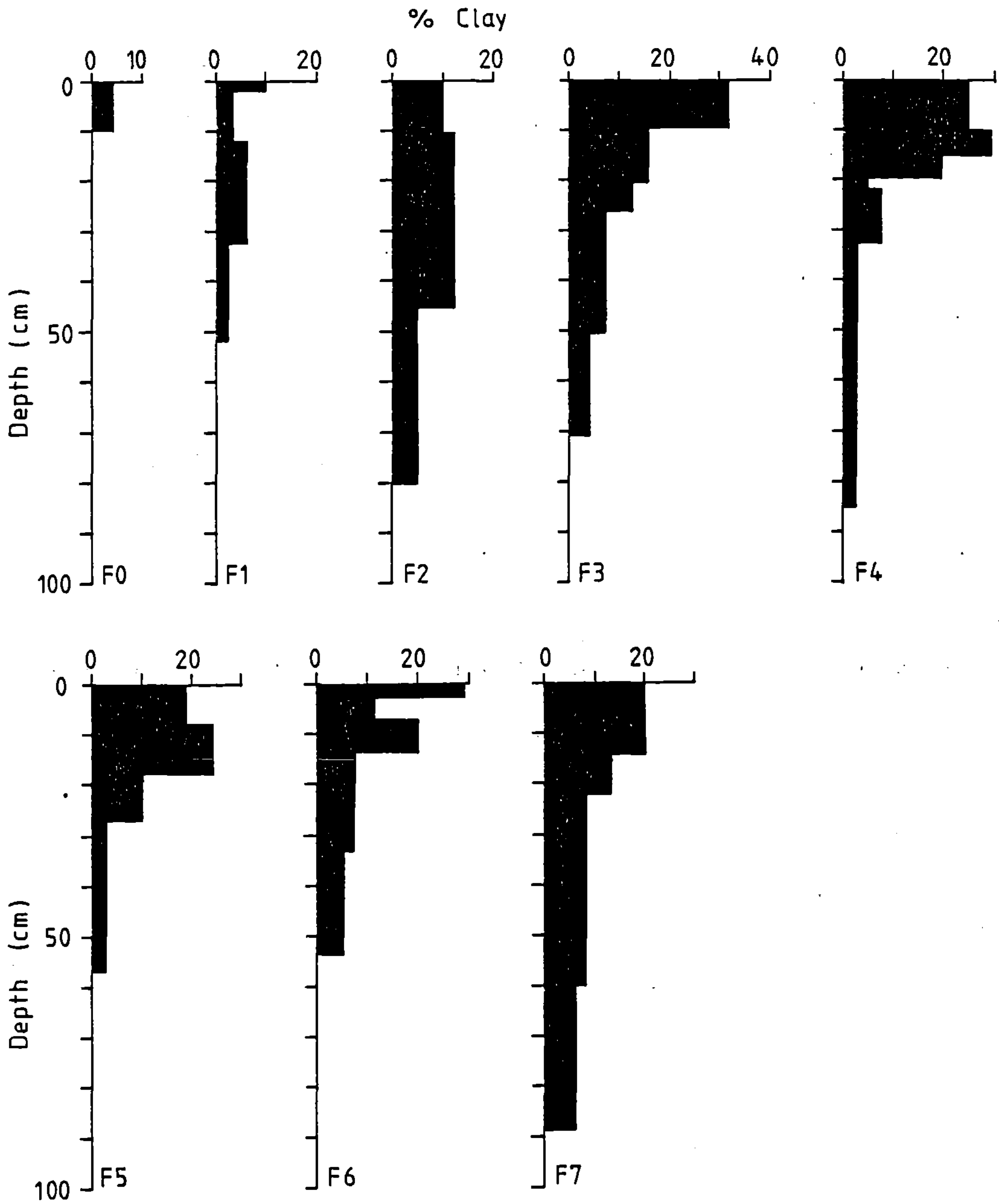


Fig. 75(a): Percent by weight of clay - soils on gentle slopes.

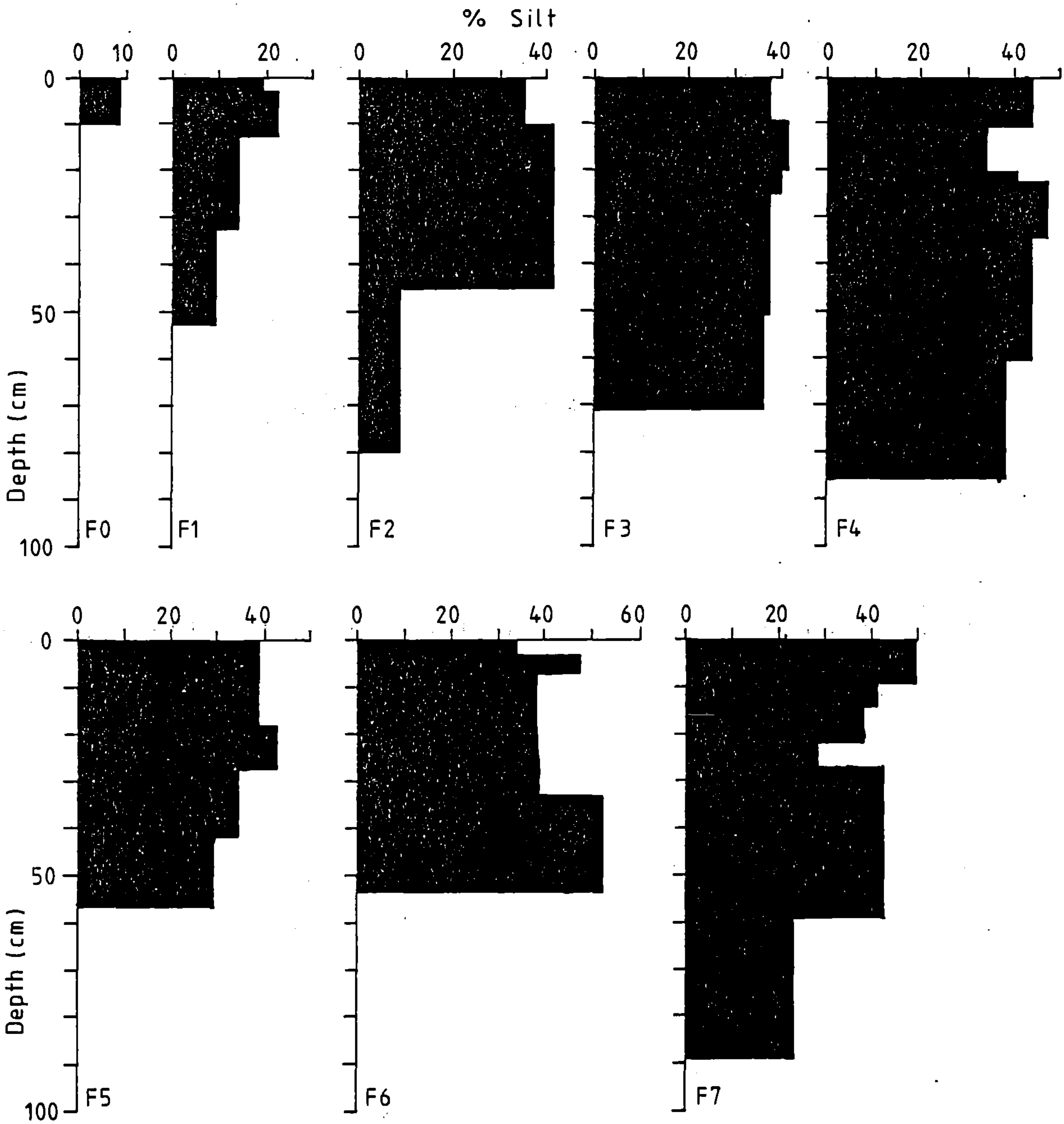


Fig. 75(b): Percent by weight of silt - soils on gentle slopes.

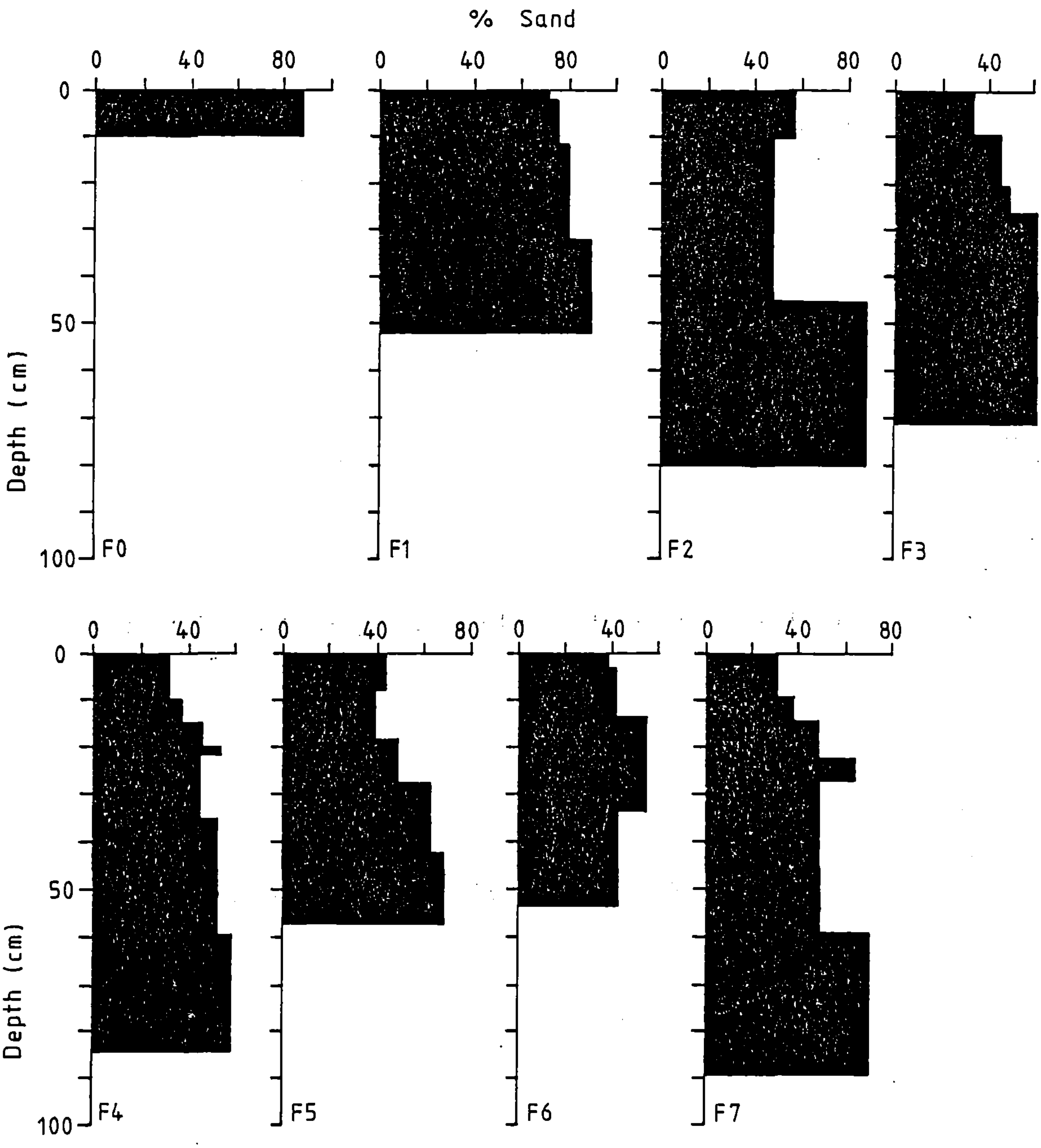


Fig. 75(c): Percent by weight of sand - soils on gentle slopes.

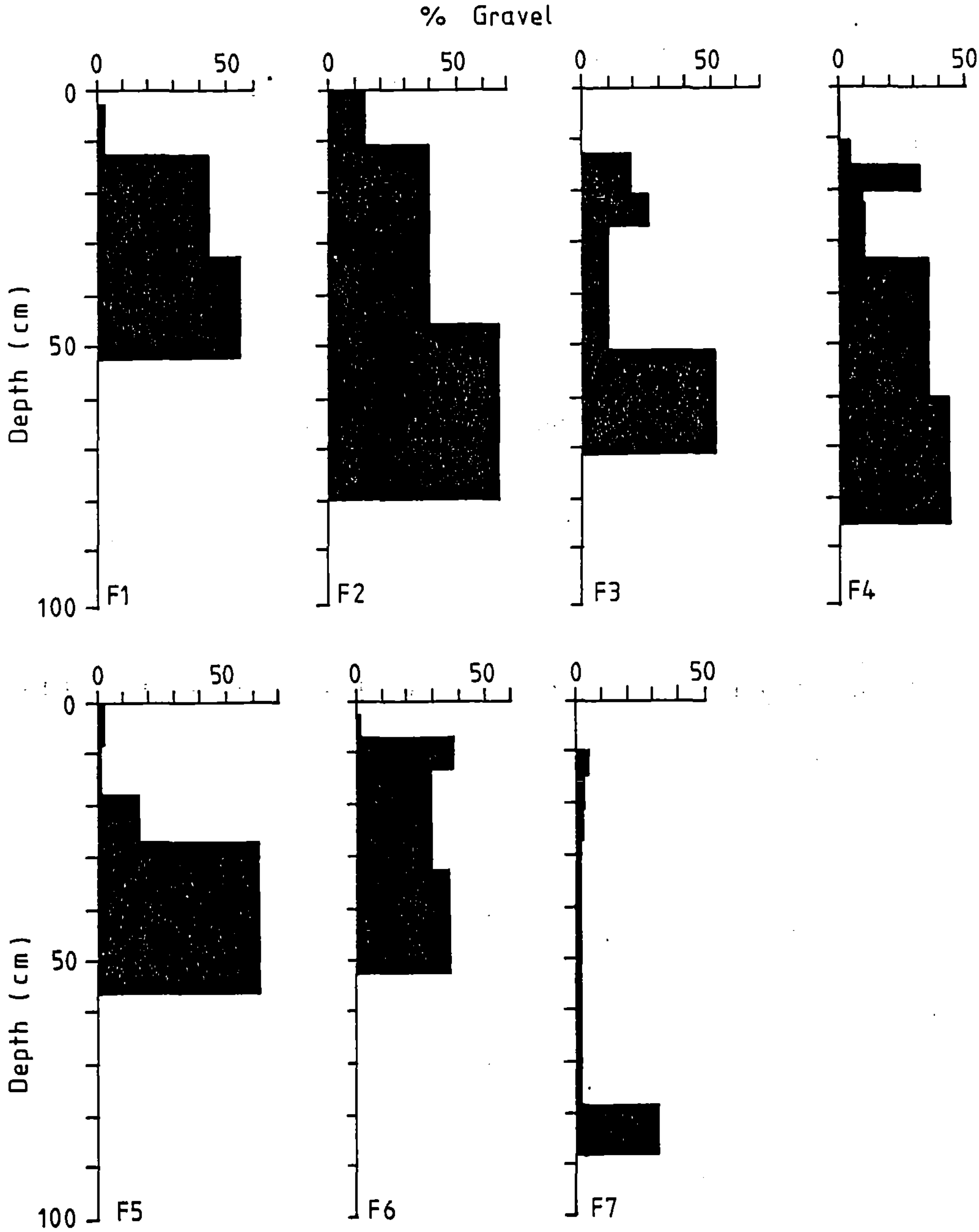


Fig. 75(d): Percent by weight of gravel - soils on gentle slopes.

variability but may also result from clay destruction at low pH in F4 and F5. Campbell (1975) also notes a decrease in clay content in the older soils of the Reefton chronosequence which he attributes to clay destruction, from exposure to organic chelates and low pH conditions over a long time period. F6 and F7 also had high clay content in surface horizons (20-29%). Clay content in F1, F2, F3 and F7 was highest in the A_h horizon, and decreased with depth, while in F4 and F5 maximum clay content was in the E horizon.

Surface horizon silt content increased from 8% in alluvium to 20% in F1 and 40% in F2. Thereafter it remained relatively constant (including F6 and F7). In the younger soils (F1, F2) silt content decreased markedly with depth, while in the older soils silt content tended to be relatively uniform or decreased slightly with depth. The high subsoil silt content is probably a result of the combined effect of silt washing in to the voids in the coarse alluvium, and *in situ* weathering. Silt caps were observed in some larger voids.

Sand content in surface horizons decreased from 88% in alluvium, to 71% in F1, 55% in F2 to 32% in F3 and F4. F5 had a slighter higher sand content (40%), although this is partly a function of the lower clay content discussed above. Sand content increased with depth in the profile and was very uniform in the alluvium, F1 and F2 at 87 to 89%. In the older soils (F3 to F5) sand content in the sampled C horizons was lower at 60 to 70% due to the relative increase in silt.

Gravel content showed similar trends to sand content i.e. a decrease in gravels in the top 20 to 30 cm of the profile with age, and an increase with depth in the profile. F1 clearly had a sandy top stratum, devoid of gravels, that was of depositional origin. F2 was gravelly throughout, although gravel content was less in the top 20 cm. The low gravel content characteristic of the topsoils of the older soils (F3 to F5) may therefore be partly contributed to by original sedimentary characteristics. However it is primarily *in situ* weathering, resulting in the breakdown of gravel and sand to produce silt and clay, that has resulted in the characteristic texture profile form of silt loam over stony sandy loam. This conclusion is based on the following:

- the older soils had gravels remaining in the topsoil that are commonly quartzose and highly weathered, particularly in the E_{ag} horizon;
- gravel content in these soils increased progressively with depth in the profile. A_h and E_{ag1} horizons had low gravel content, with increases in the E_{ag2} and $2B_s$ horizons;
- F6, the old soil formed in bedrock, showed the same characteristic of low gravel content in the topsoil.

This characteristic texture profile form can only have developed as a result of pedogenesis, or possibly by loessial input. However since these soils are all of Aranuian age, and in an area where little of the landscape is unvegetated, a significant loessial input seems unlikely.

The soils formed in coarse textured alluvium (F1 to F5) had similar gravel (45 to 65%) and clay (2 to 5%) contents in their C horizons. Sand and silt contents varied however, due to the higher silt contents in the subsoils of the older soils, as discussed previously. F6 and F7 were distinguished by their lower gravel content in the C horizon (35%).

Textural classification of the <2 mm diameter fraction of the analysed samples, using the soil texture classification of Taylor and Pohlen (1962), is illustrated in Fig. 76. Most A_h and E_{ag} horizons fell within the silt loam class; the B horizons fell in the silt loam or loamy sand class; and the C horizons and younger A_h horizons

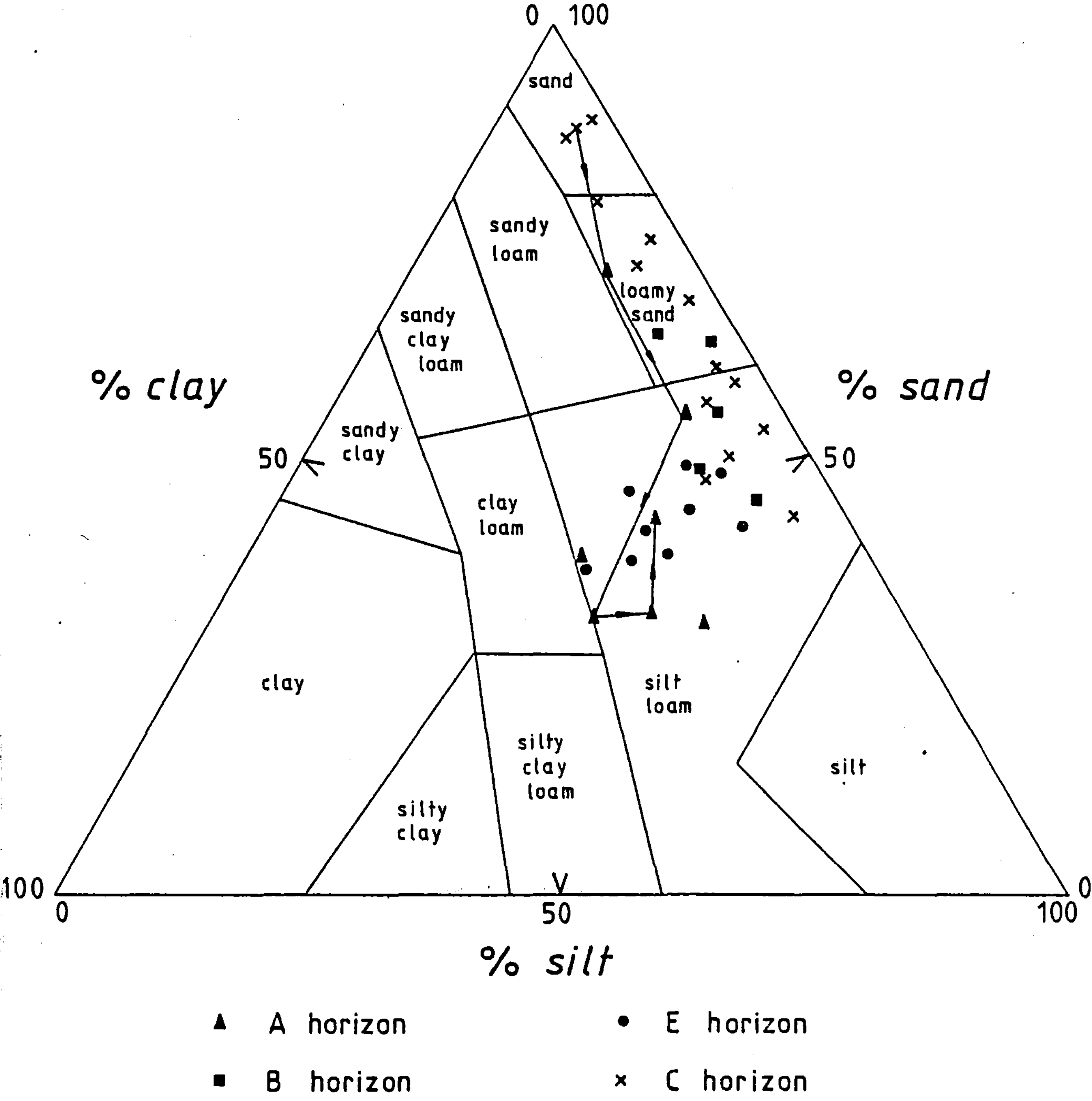


Fig. 76: Soil texture classes (fine earth fraction) - soils on gentle slopes.

fell in the loamy sand or sand class. On Fig. 76 a line has been drawn connecting the points representing alluvium (F0) and the A_h horizons of F1 to F5. This line represents the evolutionary pathway of soil development and shows an initial rapid increase in silt and decrease in sand (F0 to F2), a marked increase in clay from F2 to F3, followed by a decrease in clay from F3 to F5.

b) Soils on steep slopes (fig. 77).
Analyses for these soils show broadly similar trends to those described for soils on gently sloping sites, although the steeply sloping soils are not clearly ranked with respect to age. Gravel and sand content in surface horizons decreased with increasing soil age, while clay content initially increased (S1 to S3) and then decreased slightly (S3 to S6). Clay content decreased down the profile while gravel and sand content increased.

C horizons of these soils had uniformly low clay content (<6%), but variable gravel content. S2 and S4, formed from colluvium derived from debris avalanche material, had low gravel content (30 to 40%) as did S6 (derived from colluvium). $S3_u$, $S3_l$, $S5_u$, and $S5_l$, formed from colluvium derived from alluvium, had a 60 to 70% gravel content.

Textural classification of the <2 mm diameter fraction (Taylor and Pohlen, 1962) showed that most of the upper horizons were silt loams while the C horizons were loamy sands (Fig. 78).

6.6.9 Chemical and physical analysis of S7, S8, S9, S10.

a) Recent soils on steep slopes - Table 14
Three soils (S7, 8, 9), that were typical of the recent soils that cover the majority of the landscape, were analysed. All three were formed on steep slopes (>40°), from pelitic schist, under *Chionochloa pallens* grassland.

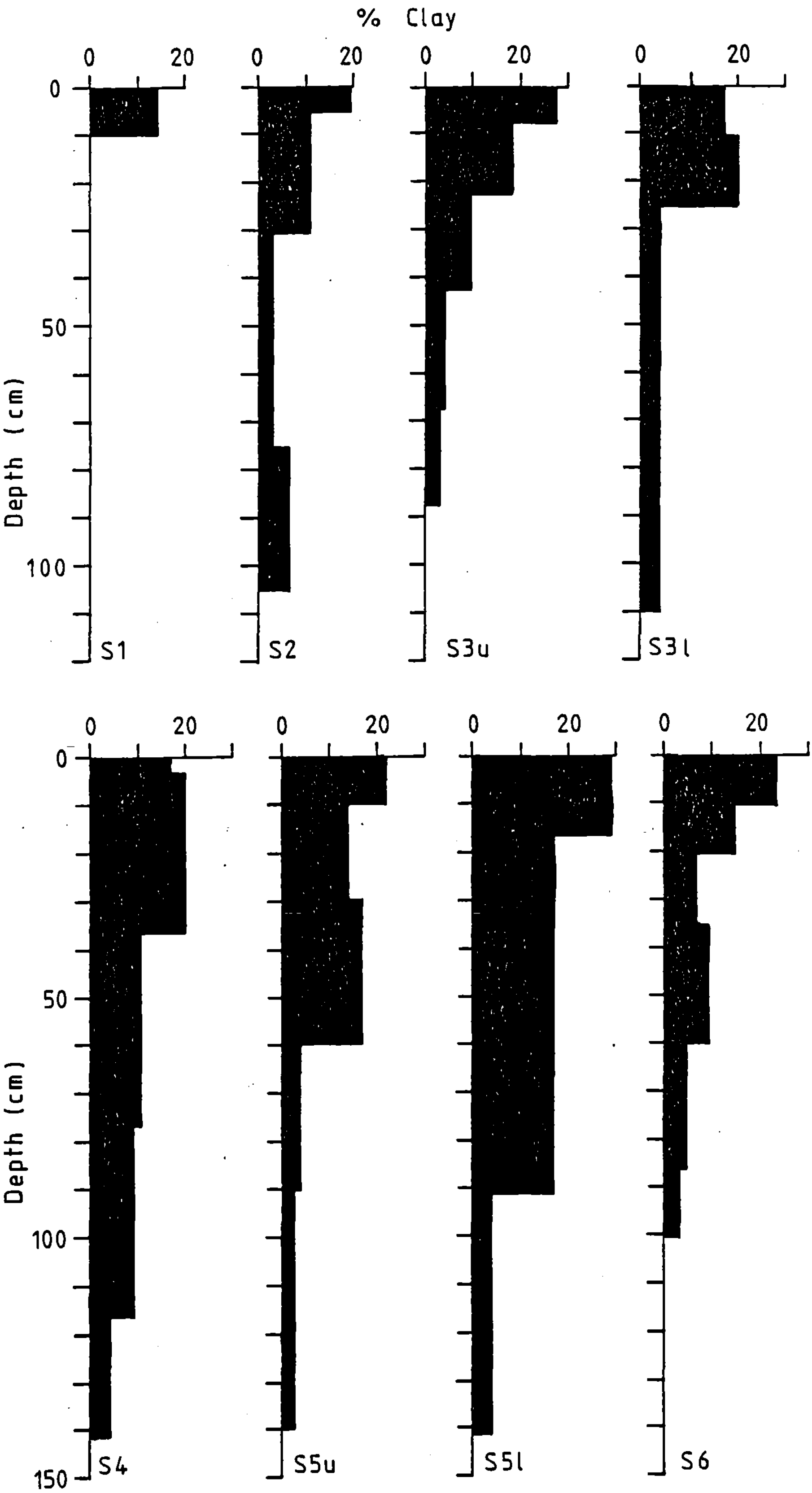


Fig. 77(a): Percent by weight of clay - soils on steep slopes.

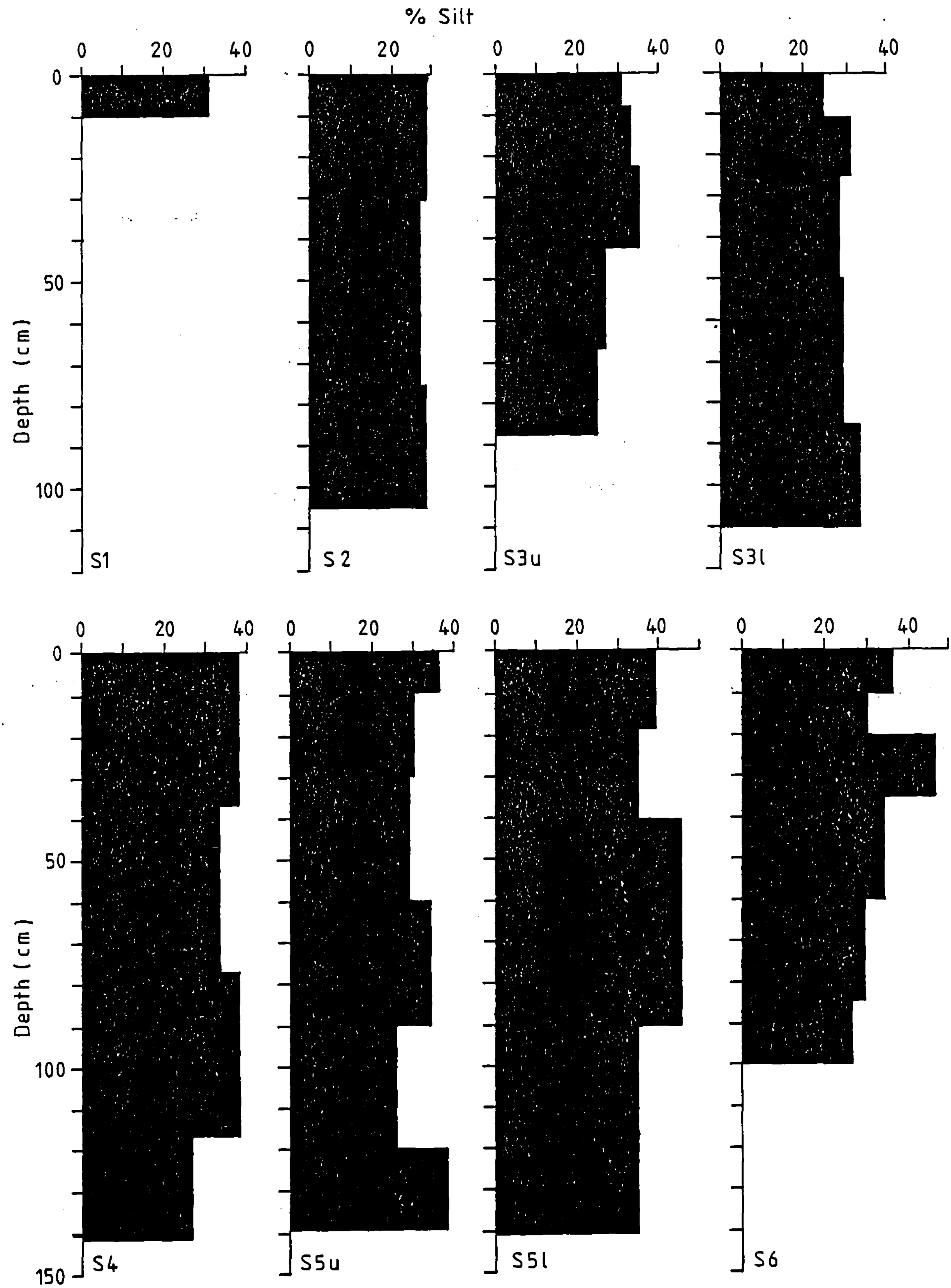


Fig. 77(b): Percent by weight of silt - soils on steep slopes.

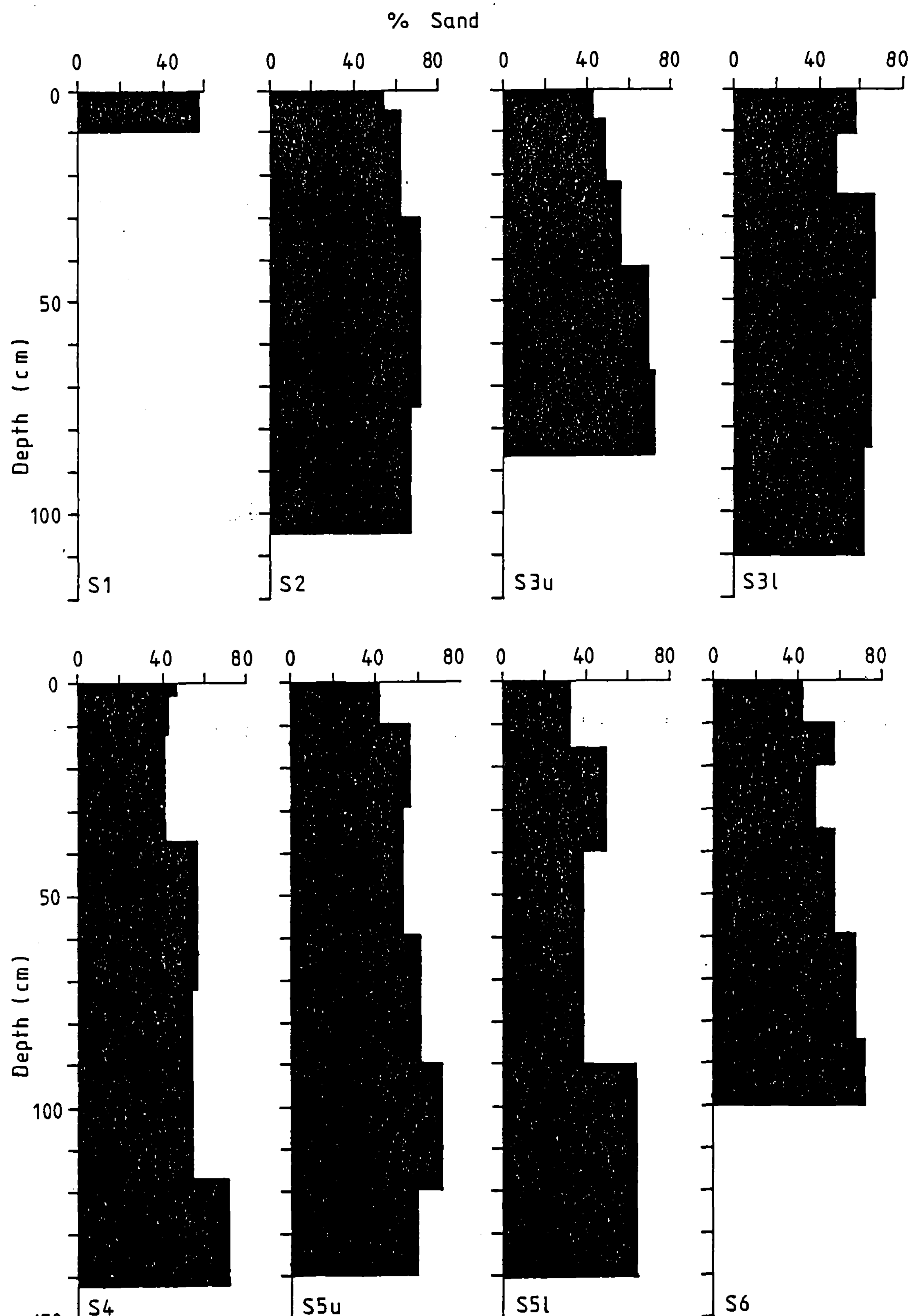


Fig. 77(c): Percent by weight of sand - soils on steep slopes.

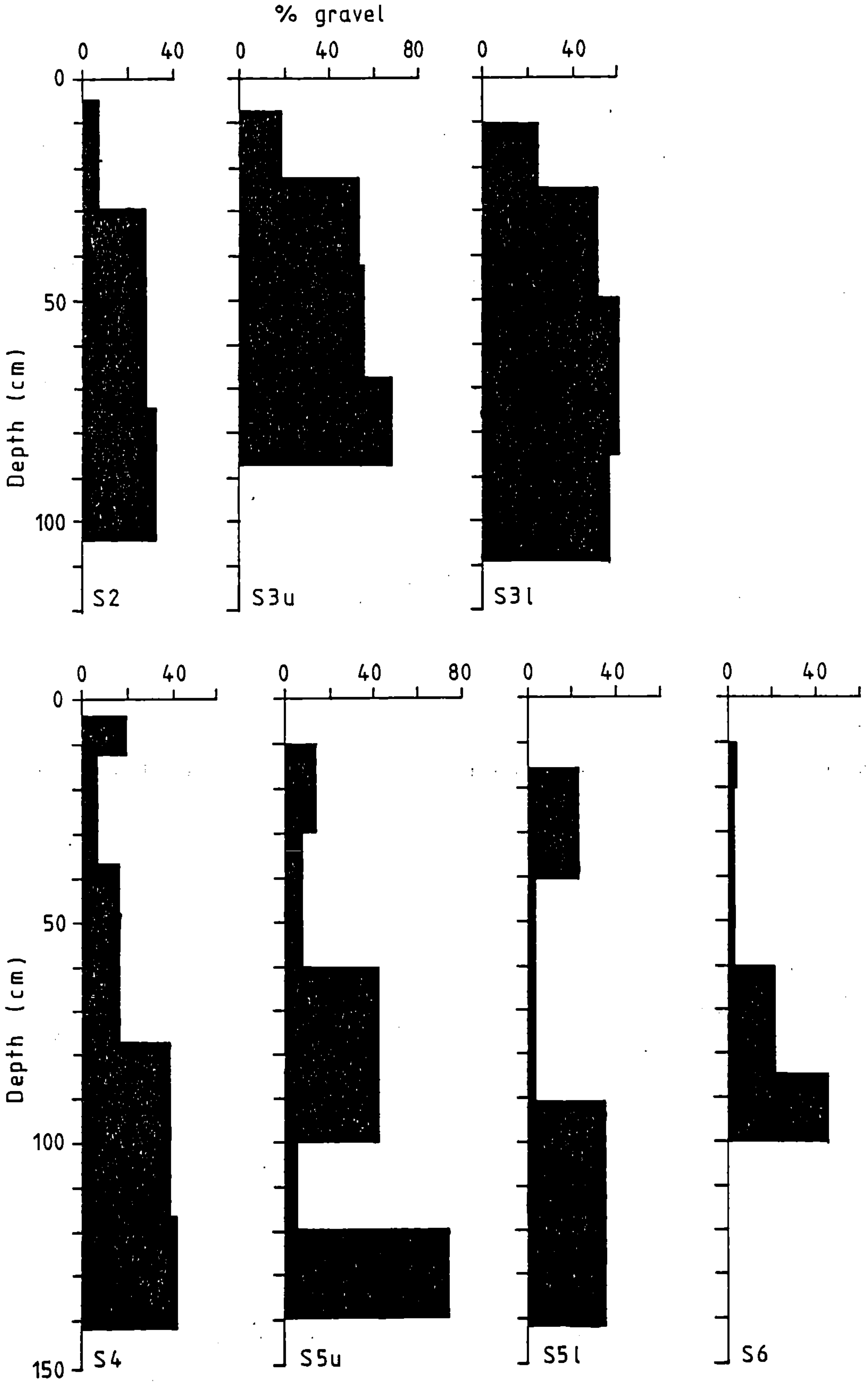


Fig. 77(d): Percent by weight of gravel - soils on steep slopes.

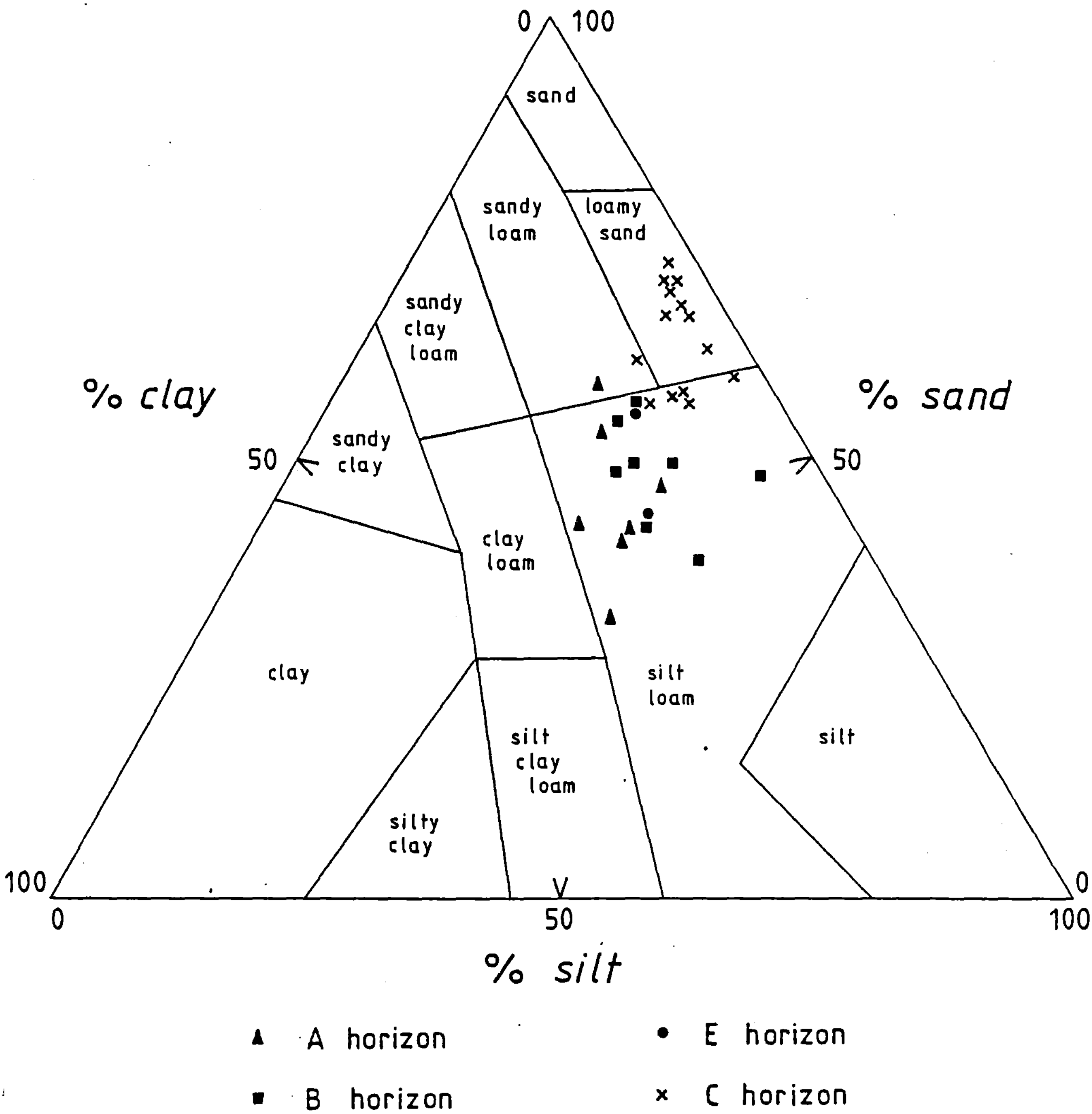


Fig. 78: Soil texture classes (fine earth fraction) - soils on steep slopes.

These soils were all strongly acid, with pH (H_2O) of the surface horizons ranging from 4.3 to 4.9. pH increased down the profile, with C horizon values ranging from 4.8 to 5.3. pH (NaF) was variable, with S7 having uniformly low values (<7.7) while S8 and S9 had a slight increase down the profile (up to 9.0). Concentrations of oxalate-extractable Fe and Al were low, with maximum values in the A_h horizons and a decrease with depth.

CEC (NH_4 OHc) was medium (14 m.eq./100 g) in the A_h horizons and decreased to low or very low values with depth. ECEC displayed the same trend, but the values were far lower than for CEC (4.4 to 5.9 m.eq./100 g). Exchangeable bases were low or very low throughout all three soils (<0.5 m.eq./100 g) as were total exchangeable bases (<1.5 m.eq./100 g). Maximum concentration in each soil occurred in the A_h horizon, and there was a decrease with depth in the profile. Exchangeable H and Al attained far higher concentrations than other exchangeable cations (up to 3.5 m.eq./100 g). They were at a maximum in the A_h horizon and decreased with depth. Base saturation (at pH 7) was very low (2 to 10%) throughout all three soils and values decreased with depth. Base saturation (at pH MKCl) gave higher values (5 to 32%) but they were still mainly low.

Particle size analysis of S7 showed an increase in gravel content with depth, while the fine earth fraction had a uniformly high sand content (55%) and low clay content (<8%).

b) Shallow soil on bedrock - Table 15

S10 is an older soil formed from resistant psammitic schist, under mixed low shrubby grassland.

Topsoil pH (H_2O) was extremely acid (4.2), rising to moderately acid in the subsoil (5.0). pH (NaF) was very low and increased slightly with depth (7.1 to 7.7).

Sample Code	Depth (cm)	Horizon	pH		LOI (%)	oxalate-extractable (%)			m.eq./100 g									Base saturation (%)	
									Exchangeable cations						ECEC	CEC	TEB		
			H_2O	NaF		Fe	Al	Si	Ca	Mg	K	Na	H	Al		NH_4OAc		at pH MKCl	at pH7
S7.1	0-10	A_h	4.7	7.4	14.7	0.43	0.13	0.010	0.3	0.5	0.46	0.15	1.1	1.9	4.4	14.0	1.4	32	10
.2	10-20	C^h	4.9	7.5	9.9	0.47	0.14	0.011	<0.05	0.2	0.20	0.09	0.9	2.1	3.5	10.5	0.5	15	5
.3	20-30	C	4.9	7.7	2.7	0.24	0.09	0.014	<0.05	<0.05	0.03	0.04	0.1	0.8	1.0	2.6	0.1	11	4
S8.1	0-5	A_h	4.9	8.0	14.6	0.91	0.26	0.035	0.1	0.4	0.45	0.15	0.8	3.1	5.0	14.6	1.1	22	8
.2	5-20	C^h	4.8	8.6	3.8	0.41	0.14	0.021	0.05	0.1	0.05	0.03	0.3	2.5	3.0	9.1	0.2	8	2
S9.1	0-5	A_h	4.3	7.9	12.3	0.58	0.21	0.034	0.2	0.4	0.36	0.13	1.3	3.5	5.9	14.1	1.1	19	8
.2	5-20	C^h	4.6	8.8	5.2	0.57	0.17	0.021	<0.05	0.1	0.10	0.04	0.6	2.7	3.6	9.2	0.3	8	3
.3	20-40	C	4.9	9.0	2.9	0.39	0.12	0.021	<0.05	<0.05	0.03	0.02	0.3	2.1	2.5	6.6	0.1	6	2
.4	40+	R	5.3	8.8	1.5	0.26	0.06	0.010	<0.05	<0.05	0.03	0.01	0.3	1.1	1.5	4.6	0.1	5	2

	Particle size analysis (% by weight)			
	gravel (>2 mm)	sand (2-.02 mm)	silt (.02-.002 mm)	clay (<.002 mm)
S7.1	25	55	37	8
.2	57	57	38	5

TABLE 14: Analyses of S7, S8 and S9.

Sample Code	Horizon	pH		oxalate-extractable (%)			LOI (%)
		H_2O	NaF	Fe	Al	Si	
S10.1	A_h	4.2	7.1	0.17	0.17	0.002	14.0
.2	E_{ag}	4.5	7.5	0.25	0.15	0.003	4.8
.3	E_{ag}	5.0	7.7	0.26	0.15	0.014	2.3

	Particle size analysis % by weight			
	gravel >2mm	sand 2-.02mm	silt .02-.002mm	clay <.002mm
S10.1	0	52	31	17
.2	7	54	30	16

TABLE 15: Analyses of S10.

Oxalate Fe and Al were very low throughout the soil. Fe_0 increased slightly with depth, while Al_0 remained uniformly low. Particle size analysis showed low gravel content (<7%) and moderate clay content (16 to 17%) in the A_h and E_{ag} horizons.

6.7 CONCLUSIONS

The results of this study are consistent with established models of soil development on quartzofeldspathic parent materials under conditions of intense leaching and acidification (Stevens, 1968; Campbell, 1975) - see 2.4. In general the analytical data support the arrangement of soils into sequences that are primarily a function of soil age.

The major trends noted with increasing soil age were:

a) A decrease in pH (H_2O) in surface horizons and establishment of a depth gradient for pH;

b) An initial increase in concentrations of oxalate-extractable Fe and Al in surface horizons, followed by a decrease in surface horizon concentrations combined with and increase in B horizon concentrations;

c) Total Ca and P concentrations in surface horizons initially declined rapidly but then changed little. Total Fe and Al concentrations in surface horizons decreased. Increases in B horizon concentrations of total Fe occurred in the older soils. Si increased with time as Si was less mobile than Fe or Al.

d) Zr accumulated in surface horizons with increasing soil age and was used to assess parent material uniformity and the degree of soil development (expressed as eluvial-illuvial coefficients). Most coefficients were negative indicating loss of Ca, P, Fe, Al and Si. In some surface horizons coefficients for Ca and P were positive, and in some illuvial horizons coefficients for Fe, Al, Si were positive. Losses of all elements from the surface soil horizons tended to increase with time;

e) There was an overall decrease in % sand and gravel and a proportionate increase in % silt and clay in surface horizons.

These trends were most clearly expressed in soils on gently sloping sites formed from coarse textured alluvium (F1 to F5). Analyses of total element concentrations, Zr concentration, and ratios of Si/Fe, Si/Al, Al/Fe indicated a high degree of parent material uniformity for these soils. Analyses for F6 and F7 confirmed their interpreted age. F6 (an old soil formed from bedrock) was morphologically less developed than F5 or F4, but chemically highly weathered. F7 (a younger soil from fine textured alluvium) was morphologically strongly developed, but intermediate in chemical weathering between F3 and F4. Soils on steep slopes showed the same trends (a to e above), but the soils were not always clearly ranked with respect to age. C horizons of these colluvial soils were more variable in both physical and chemical composition than soils from alluvium.

Three pairs of soils, each on surfaces of the same age were compared to assess the effect of slope position on soil chemistry. The most consistent trends noted were that soils in topographically lower positions had lower topsoil pH (H_2O) and lower concentration of both oxalate-extractable and total Fe. These trends were confirmed by the E-I coefficients of total Fe which showed greater loss of Fe in lower slope positions. These relationships were more strongly expressed in the older pairs of soils (F4 and F4p, $S5_u$ and $S5_l$) than the younger pair ($S3_u$ and $S3_l$). Al and P also tended to accumulate in soils in lower slope positions associated with the development of peaty surface horizons. The lack of strongly expressed catenary relationships is consistent with the suggestion by Tonkin (1984) that classical soil catenary relationships (in terms of chemical differentiation) are a transitory phase attained at an intermediate stage of leaching. With intense leaching soils attain a "uniformly" leached state in all slope positions and catenary relationships are only expressed in the least mobile constituents such as Fe and Al.

The use of eluvial-illuvial coefficents for evaluating profile development was tested on a wide range of soils. These coefficients were useful for differentiating soils with significant age differences (e.g. F1 c.f. F3 c.f. F4) but were less able to distinguish soils that were not greatly different in age (e.g. F1 and F2, F4 and F5). They also ranked the soils on gently sloping sites more clearly with respect to age than the soils on steep slopes, primarily because of the greater parent material uniformity of alluvium compared with colluvium. Rock analyses indicated significant variation in chemical composition of the major bedrock lithologies. Fluvial transport and deposition probably sorted bedrock lithologies more effectively than mass movement processes. Theoretically one would expect fluvial sorting to enrich the alluvium in the more resistant psammitic rocks. These had less of a range in elemental composition than the pelitic rocks that were analysed. An assumption in calculating eluvial-illuvial coefficients was that the C horizon was nearest in composition to that of the original material in which a soil had developed. There were indications that the depth of weathering had exceeded the sampling depth in the older soils, and therefore the degree of soil development was underestimated. This may be partly responsible for the inconsistent ranking of some of the soils on steep slopes (S4, $S5_u$, $S5_l$). Total Zr was suitable as an internal standard, being the least mobile of the elements determined. It was not confirmed that Zr remained unaffected by pedogenesis, therefore estimates of change relative to Zr are minimum estimates.

Soil fertility characteristics were determined for a young (recent) soil and an old (gley podzol) soil on both gently and steeply sloping sites. Analytical data showed:

a) Exchangeable cations concentrations were low in both young and old soils, with a predominance of H and Al on exchange sites. Base saturation was also extremely low;
b) Low concentrations of H_2SO_4-soluble P in all surface, horizons, and an increase with depth in the profile. The opposite trend was noted in organic P i.e. high

concentration in surface horizons and a decrease with depth. Even the two young soils had a low ratio of 0.5M H_2SO_4-soluble P to total P in surface horizons.

These characteristics reflected the strong leaching and weathering regime of the high rainfall environment. Phosphorus analyses also suggested differences in parent material composition between alluvium and colluvium. The colluvium had lower levels of total P and 0.5 H_2SO_4-soluble P a lower ratio of 0.5M H_2SO_4-soluble P to total P. This results from the pre-weathered nature of the colluvial parent materials.